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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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FISSILITY OF MUDROCKS

By Roy L. Ingram

Abstract

The breaking characteristics of a mudrock can be represented on a triangular diagram with massive, flaky-fissile, and flaggy-fissile as end members. Fissility in shales is usually associated with a parallel arrangement of the micaceous clay particles and nonfissility with a random arrangement. Experiments and observations indicate that the clay minerals attain a parallel arrangement by gravity settling or by flocculation and compaction, unless the particles are adsorbed on irregularly shaped sesquioxide or silica particles or grow randomly in a gel. The nature of the cementing agent determines whether a shale will be flaky or flaggy. If the cementing agent can hold the material in a large slab, the shale will be flaggy. If the amount or the tenacity of the cementing agent is small, the shale will be flaky. Cementing agents other than organic matter tend to hinder cleavage parallel to the clay particles causing a decrease in fissility and an increase in massiveness. Moderate weathering increases the fissility of a shale. In general the type of fissility does not correlate with the type of clay minerals present in a random collection of mudrocks.

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Introduction

The purpose of this study is to explain the origin of fissility and nonfissility in nonlaminated mudrocks. Fissility caused by separation along planes between beds or laminae of different lithologic characteristics is outside the scope of this paper.

The literature shows that the terminology used in describing argillaceous sediments is not standardized (Lewis, 1924; Bradley, 1931a; Twenhofel, 1937; Krynine, 1948; Dapples, Krumbein, and Sloss, 1950). Therefore it is necessary to define certain terms as used in this paper.

Fissile—possessing the property of splitting along approximately parallel surfaces

Massive—nonfissile

Mudrock—sedimentary rock of which at least 50 per cent is silt and clay, with no connotation as to the relative percentage of silt and clay and as to the breaking characteristics

Siltrock—mudrock with silt dominant over clay

Clayrock—mudrock with clay dominant over silt
Mudstone—massive mudrock
Siltstone—massive siltrock
Claystone—massive clayrock
Mud shale—fissile mudrock
Silt shale—fissile siltrock
Clay shale—fissile clayrock

Table 1 summarizes this scheme.

TABLE 1.—NOMENCLATURE OF SEDIMENTARY ROCKS CONTAINING MORE THAN 50 PER CENT SILT AND/OR CLAY

	No connotation as to breaking characteristics	Massive	Fissile
No connotation as to relative amounts of silt and clay	Mudrock	Mudstone	Mud shale
Silt predominant over clay	Siltrock	Siltstone	Silt shale
Clay predominant over silt	Clayrock	Claystone	Clay shale

Mudrocks of many varieties were collected. Analyses were made in hopes of finding parameters that vary with fissility. Attempts were made to reproduce experimentally fissile and nonfissile mudrocks. Finally, post-depositional changes that might influence fissility were considered.

REVIEW OF THE LITERATURE

A perusal of the literature pertaining to the structure of mudrocks reveals: (1) the emphasis on the explanation of fissility whereas nonfissility apparently has been assumed to need no explanation, and (2) the large amount of pertinent work by the soil scientists which has not been assimilated into the geologic literature.

Ideas expressed in geologic literature as to the cause or partial cause of fissility are: (1) rotation of flaky clay particles into positions perpendicular to the force exerted by the overlying material with concurrent flattening of clay aggregates (Grabau, 1913, p. 175; Lewis, 1924, p. 520, 578; Harker, 1932, p. 157–158; Hedberg, 1936, p. 269; Hatch and Rastall, 1938, p. 122; Twenhofel, 1939, p. 293–294; Milner, 1940, p. 386–387; Alling, 1946, p. 26; Pettijohn, 1949, p. 288); (2) settling of flaky particles in water with the flat surfaces lining up parallel to the surface of deposition (Mohr, 1932, p. 130; Kerr, 1937, p. 545; Keller, 1946, p. 71; Krumbein, 1947, p. 102); (3) orientation of clay particles by weak currents (Keller, 1946, p. 71); (4) penecontemporaneous and epigenetic growth of clay minerals parallel to the bedding (Lewis, 1924, p. 578; Keller, 1946, p. 70; Twenhofel, 1939, p. 293–294; Payne, 1942, p. 1717; Pettijohn, 1949, p. 278); (5) presence of illite (Grim and Cuthbert, 1945, p. 94; Keller, 1946, p. 71); (6) parting caused by expansion after the removal of overburden by erosion (Hedberg, 1936, p. 245); (7) low carbonate content (Campbell, 1946, p. 832); (8) high sulfur and carbon content (Campbell, 1946, p. 832); (9) weathering phenomena (Bain, 1896, p. 293; McKelvey, 1946, p. 28–29); (10) necessary small grain size of the nonclay minerals (Alling, 1945, p. 739); (11) minute stratification (Grout, 1932, p. 269; Twenhofel, 1939, p. 293–294). Bradley (1931b, p. 319) and Grout (1932, p. 269) were intentionally noncommittal as to the origin of fissility in many mudrocks.

Nonfissility has been explained by: (1) deposition of a large quantity of colloidal clay gel, thereby preventing the reorientation of randomly oriented clay particles (Lewis, 1924, p. 580); (2) random growth of clay minerals in a clay gel (Keller, 1946, p. 69–70); (3) stirring by wave action (Weller, 1930, p. 129); (4) consolidation of massive clay soils (Keller, 1946, p. 71); (5) wind deposition of argillaceous material (Keller, 1946, p. 71); (6) presence of iron oxides in clay (Keller, 1936, p. 55; Alling, 1945, p. 740); (7) deficiency of potassium (Lewis, 1924, p. 581); and (8) presence of excessive nonclay size, nonclay mineral materials (Alling, 1946, p. 18, 25, 35). Pettijohn (1949, p. 280) presented no explanation for the massiveness of argillites even though they do not "differ in any important respects from normal shales and slates."

EMPIRICAL CLASSIFICATION OF FISSILITY

Mudrocks were collected from the Eocene of Colorado; the Cretaceous of Iowa and Colorado; the Jurassic, Triassic, and Permian of

Colorado; the Pennsylvanian of Iowa and Colorado; the Mississippian of Iowa; and the Ordovician of Wisconsin. Three dominant types of breaking characteristics were observed: massive, flaggy-fissile, and flaky-fissile (Pl. 1).

Massive mudrocks have no preferred direction of cleaving or breaking. Most of the fragments are blocky. A few of the fragments may be platy, but the orientation of the platy fragments in the outcrop is random.

Flaggy shales split into fragments of varying thicknesses but with the width and the length many times greater than the thickness and with the two essentially flat sides approximately parallel. Most of the fragments in the flaggy shales have the length and width at least 50 times greater than the thickness. Some of the fragments may break so as to fit the description of the flaky shales, but these are not so dominant as the flaggy fragments.

Flaky shales split along irregular surfaces parallel to the bedding into uneven flakes, thin chips, and wedgelike fragments, whose length seldom exceeds 3 inches.

Some samples combine varying amounts of massiveness, flagginess, and flakiness, producing a continuous coverage of a triangular diagram if the three dominant types are placed at the apices of the triangle (Fig. 1). This classification is only qualitative, but a given mudrock can be located easily if the end members are kept in mind. This was verified by the close agreement in the location of a number of samples on the triangular diagram by several persons.

All attempted verbal descriptions for the breaking characteristics of all possible combinations of massiveness, flagginess, and flakiness were too unwieldy. Mudrocks with intermediate characteristics can be described as dominantly massive, dominantly flaggy, or dominantly flaky.

Mudrocks with splintery breaking characteristics do exist. The splintery nature is most probably caused by closely spaced fractures and is not considered in this paper.

Analyses and Studies of Mudrocks

Color

Attempts to correlate color with type of fissility yielded the following observations: (1) those with a flaggy structure are predominantly black or dark gray, (2) those with a flaky structure are predominantly gray or gray black and, in general, are not so dark as the flaggy ones, (3) those with a massive structure have a predominance of colors other than gray black or black—*i.e.*, gray, white, yellow, and red, (4) some exceptions exist to the above trends, the most notable being the occurrence of predominantly flaggy red mudrocks.

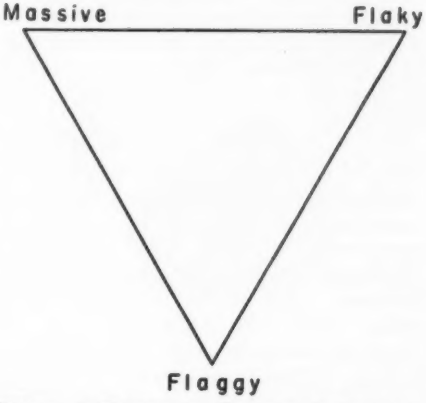

Figure 1. Triangular Diagram Showing a Megascopic Classification of Mudrock-Breaking Types

Mechanical Analyses

The dispersing and centrifuging procedure used for mechanical analyses was evolved by Jackson, Truog, and their students in the Soils Department of the University of Wisconsin (White and Jackson, 1946, p. 150–154). This procedure would have been unnecessary if only mechanical analyses were desired; but, since separations were necessary to obtain samples for x-raying, this procedure was most expedient.

Fifty samples, representative of the variations in fissility were chosen for more detailed studies. Each sample was split into four size groups: sand and gravel (greater than 20 microns), silt (2–20 microns), coarse clay (0.2–2 microns), and fine clay (less than 0.2 microns).

No correlation exists between the results of the mechanical analyses and the type of fissility. This is expected since only a small amount of clay is necessary to control the behavior of a sediment (Nutting, 1927, p. 185; Sideri, 1938,

p. 130; Alling, 1946, p. 26), and the addition beyond the minimum required should have little effect.

The results of a mechanical analysis of a mudrock have little meaning since they depend on so many variables. Clay minerals vary widely in their ability to be dispersed (Hauser and LeBeau, 1946, p. 202). Different dispersing ions cause different degrees of disaggregation (Gripenberg, 1939, p. 532; Grim, 1939, p. 488; Hauser and LeBeau, 1946, p. 203). Sideri (1936, p. 461, 465) has shown that clay, organic matter, and fine sand form a system that is impossible to disperse completely. The smallest grains are always aggregates. In experiments performed by the writer even 15 to 20 treatments with 30 per cent H_2O_2 had little effect in removing organic matter from some black shales. Microscopic studies show that the mudrocks contain less sand and silt than the analyses indicated. Even if one assumes that complete dispersal could be attained, it does not follow that this duplicates the conditions of deposition of the sediments.

Alling (1945, p. 739, 751) has shown that an increase in silt and sand causes a decrease in fissility. This relationship was not investigated in this work. But this is not the only parameter of fissility since some clayrocks are fissile and some are massive.

Carbonate Content

An estimate of the carbonate content was obtained by adding a standardized acid in excess to a powdered sample and back titrating with a standardized base using a phenolphthalein indicator (Willard and Furman, 1940, p. 136–152). The acid consumed was assumed to have been used in reacting with $CaCO_3$. For these 50 samples no correlation exists between the amount of $CaCO_3$ and the type of fissility.

Peterson (1944, p. 44) observed a decrease in the platiness of a dried puddled kaolinitic soil upon the addition of $Ca(OH)_2$. This increase in blockiness or granularity is attributed to the cementing effect of $Ca(OH)_2$. Campbell (1946, p. 832) observed a similar trend in the New Albany shale, where the fissility varied indirectly with the lime content. Undoubtedly the cementing action of carbonates causes a decrease in fissility with a given type of shale. Because so many other factors influence fissility, it is easy to see why the carbonate content of a random collection of mudrocks does not vary with fissility.

Organic Content

The organic content of the mudrocks was estimated by the Schollenberger chromate titration method (Schollenberger, 1931, p. 483–486). The organic content is high for flaggy shales, moderate for most flaky shales, and low for most massive mudrocks. These results are discussed later in the paper.

Clay-Mineral Studies

The techniques used for the preparation, x-raying, and interpretation of the x-ray diffraction data are those developed by Jackson and his students (White and Jackson, 1946, p. 150–154). Each clay was saturated with calcium ions and glycerol solvated. Unfiltered iron radiation was used on wedge-shaped samples in a camera with a diameter of 143.2 mm.

Montmorillonoid, chlorite, illite, kaolinite, and mica intermediates were identified. No correlation exists between the presence or amount of any clay mineral and the type of fissility in the 50 samples studied. The clay minerals identified are distributed among all the types of fissility.

Several workers have expressed the idea that the type of clay mineral determines whether or not a mudrock will be fissile. Payne (1942, p. 1717) and Grim and Cuthbert (1945, p. 94) believe that sediments composed of illite will have a greater tendency toward fissility than those containing other clay minerals because illite will not break into individual layers in the presence of water as will montmorillonoid and kaolinite. Marshall (1936, p. 29) expects a stronger tendency toward oriented coagulation with those clay minerals, illite and beidillite, that possess the strongest surface charge as a stronger pull will be exerted on the adsorbed cations.

Some observations and experiments show the kinds of clay minerals in fissile and massive mudrocks. Kerr (1937, p. 545) described the fissile Attapulgus, Georgia, clay as containing

FIGURE 1. MASSIVE

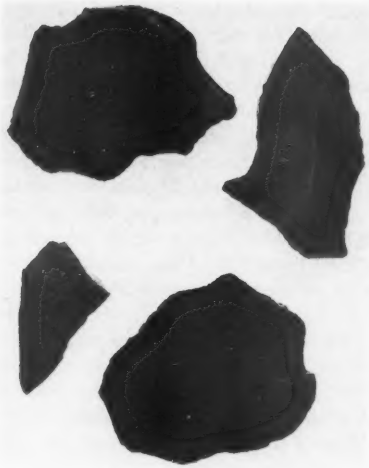

FIGURE 2. FLAKY

FIGURE 3. FLAGGY

The flaggy specimens were broken from larger slabs.

END MEMBERS OF MUDROCK-BREAKING TYPES. × ¾.

parallel montmorillonite flakes. Keller (1946, p. 69–70) observed that the massive Cheltenham fireclay of Missouri contained randomly oriented montmorillonite and illite. Peterson (1944, p. 37, 41, 44) puddled clay and fine sand, subjected the mixture to ten cycles of alternate wetting and drying, and studied the resulting product megascopically and microscopically. With the clay being either pure kaolinite or montmorillonite a massive structure was obtained, but with the addition of 3 per cent of montmorillonite to kaolinite a platy structure resulted. These observations and experiments show that montmorillonitic mudrocks may be either fissile or massive, yet Peterson's work implies that the type of clay mineral affects the breaking characteristics.

None of the samples studied by the writer contained halloysite, but mudrocks composed of halloysite are probably massive. Brindley, Robinson, and Goodyear (1948, p. 427) found natural halloysites to have a random arrangement of the halloysite particles. They obtained partial orientation by pressures of 12,000 to 16,000 pounds per square inch and state, "this, we believe, is the first time such an orientation effect has been detected with halloysite." Their electron-microscope photographs of sedimented halloysite revealed no orientation of the lath-shaped crystals.

Thus, in a random collection of mudrocks there is no apparent correlation between fissility and the type of clay mineral. The writer's success in producing fissile and massive mudrocks experimentally using the same clay mixture is more evidence against a dominating influence of different clay minerals. This does not eliminate the possibility that clay minerals influence fissility, for other parameters of fissility may mask clay-mineral effects, and further research in clay mineralogy may subdivide the known clay minerals into more species.

Microscopic Studies

Thin sections were studied to determine the microstructural features of fissile and non-fissile mudrocks. The clay minerals provide the main structural control for mudrocks. Even though most of the clay particles are too small to be seen with the microscope, the mass orientation can be determined since extinction is parallel or approximately parallel to the cleavage (Grim, 1939, p. 470; Ross, 1945, p. 181). Mudrocks with the clay-mineral flakes parallel give extinction under crossed nicols when the direction of light vibration parallels the long direction of the clay-mineral flakes. With a random orientation of the clay particles, the birefringence of the clay matrix does not change with rotation of the stage. The degree of preferred orientation can be estimated from the change of birefringence with rotation of the stage.

The mudrocks studied are composed primarily of a clay or organic matter matrix with embedded grains of detrital minerals, shells, and shell fragments and some pigments of iron oxides and organic matter. Clay is present as: (1) flaky particles or aggregates with extinction parallel to the long direction of the flakes, (2) flaky particles with admixed calcite, the admixed calcite making it impossible to determine the extinction direction of the clay, (3) radial aggregates, and (4) a matrix of clay particles too small to be seen individually. The clay particles in the matrix can have preferred or random orientation. Organic matter occurs as: (1) opaque fragments that are either equidimensional or linear in cross section and (2) brownish or blackish organic stains. Quartz and other detrital minerals are present as angular grains of fine sand, silt, and clay. The quartz grains are evenly disseminated throughout the clay matrix except for a few specimens which contain laminae of fine sand and silt. Calcite is present as: (1) scattered grains and (2) concentrated patches. Many of the sections cut across fossils. Iron oxides occur as fine dust, covering or staining the other materials.

The fissility of a mudrock depends primarily on the fabric of the clay minerals. All the fissile mudrocks exhibit some degree of parallelism of the flaky constituents; the degree of fissility in general, is proportional to the degree of preferred orientation. All the flaggy shales studied possess a lenticular type of parallelism; clear lenses of clay or clay and calcite are embedded in an opaque organic matrix. Especially notable is the tendency for the organic stain to be associated with clays that show preferred orientation. One sample is laminated by the alternation of layers with and without an organic stain.

The clay particles in the organic-stained laminae have an orientation parallel to the lamination; those in the unstained organic layers have a random orientation. The flaggy shales have appreciably more organic matter than the flaky shales. Apparently the cementing action of the organic matter aids in causing the rocks to break into large slabs. Massive mudrocks have a random orientation of the platy clay particles.

Experimental Reproduction of Fissilities

Gravity Settling

Different types of clay systems were dried to determine if different types of fissilities could be reproduced experimentally.

The clay suspensions obtained by centrifuging during mechanical analysis were combined according to color to get large volumes of gray, black, and red suspensions. These suspensions were used in all the succeeding experiments.

To duplicate these experiments, the following procedure should be used: (1) fill with clay suspension a 2 by 15 cm. soft test tube that has been flattened on the bottom, (2) evaporate to dryness in a 40°C. oven, (3) pour a mixture of Canada balsam and xylol on the dried clay, (4) put in vacuum for 2 to 10 hours, (5) cure in a 55°C. oven, (6) make a thin section and examine with a microscope.

The gray clay had a strong preferred orientation of the clay particles parallel to the bottom of the tube. Incipient horizontal cracks were present. The black clay had a strong preferred orientation parallel to the bottom with fine laminae of black organic matter. The red clay had a random orientation of the clay minerals but with a slight lamination caused by the concentration of iron oxide in layers. Horizontal partings tended to occur along these laminae. Apparently red fissile shales with a random orientation of the clay minerals could be formed in this way, but none were observed in the samples studied.

Flocculation

Clay suspensions were flocculated with electrolytes under a variety of conditions to determine if the chemical composition of the clay system influences fissility. The procedure in these experiments was essentially the same as for gravity settling except that the clays were flocculated with various electrolytes, allowed to settle, and then dried and impregnated.

Sodium chloride was added to gray, black, and red clay suspensions until they flocculated. The gray and black clays had a pronounced parallel orientation of the clay minerals and a tendency toward parting parallel to the bottom of the test tubes. The red clay showed a preferred orientation in a few spots, but most of the section exhibited a random orientation.

Flocculation of a gray-clay suspension with $FeCl_3$ gave a red mudrock with a parallel orientation. This can explain why some red mudrocks may be fissile and have a parallel structure. A similar flocculation with $AlCl_3$ also gave a parallel orientation.

To determine the influence of a gelatinous precipitate of $Fe(OH)_3$ on the structure of a mudrock, NH_4OH was added to a gray-clay suspension containing $FeCl_3$ in solution until both the clay and $Fe(OH)_3$ flocculated. The resulting red mudrock had a random orientation of the clay particles. A similar experiment in which $Al(OH)_3$ gel was produced gave a mudrock with a parallel orientation of the clay particles.

When a gray-clay suspension containing Na_2SiO_3 was flocculated with HCl, a stiff silica gel was formed with the clay dispersed throughout the gel. Upon drying, a hard chertlike mass was formed. The basal portion contained a concentration of clay that was strongly oriented parallel to the bottom. The rest of the clay had a random orientation.

Powdered Fe_2O_3, Al_2O_3, and SiO_2, obtained by drying gels of $Fe(OH)_3$, $Al(OH)_3$, and $Si(OH)_4$, were added separately to three gray-clay suspensions. Flocculation of these suspensions produced mudrocks with random orientations of the clay particles.

Discussion of Results

Gravity settling.—Gravity settling or evaporation of a dispersed clay results in a parallel arrangement of the flaky clay particles. Hauser (1939, p. 214) described this disposition as the one which gives the minimum free energy to

the system or, as stated in a later paper (Hauser, 1941, p. 188), "charged particles, if forced to settle, have the ability to pack as closely together as possible." The hydrosphere surrounding each clay particle acts as a lubricant permitting adjustment of the particles.

Hauser has done much work on the formation of clay films by the evaporation of clay suspensions. Careful evaporation of an extremely fine bentonite suspension resulted in gelation at a certain concentration (Hauser, 1939, p. 224). Ultramicroscopic examination showed that there was no geometrical arrangement of the particles. Continued evaporation resulted in an alignment of the particles into larger "crystals" (Hauser and LeBeau, 1938, p. 965) with individual particles aggregated into threads and filaments which maintained a crystallographic continuity.

Other workers have obtained similar results. Bray, Grim, and Kerr (1935, p. 11) and Grim (1934, p. 45–47) were able to make optical measurements on clay by allowing it to settle on a glass slide, the particles settling on one another with their c axes parallel. Kerr (1937, p. 545) obtained a parallel arrangement by the settling of a montmorillonitic clay. White and Jackson (1946, p. 152) found that settling of clay in water or any other dipolar liquid resulted in a parallel arrangement of the clay particles. Chang (1941, p. 220–221), in working on the structure of adobe, came to the same conclusion concerning a parallel arrangement resulting from gravity settling of dispersed clay.

Flocculation.—Coagulation of a dispersed clay by the addition of an electrolyte results in the formation of a gel or a gelatinous precipitate with the clay particles coalescing in a random manner (Hauser, 1939, p. 212–213). This gives a loose packing and a high porosity. Because of the high porosity of flocculated clay, it should be expected that compaction or drying will result in the flaky particles being oriented parallel to the bottom of the container. This explains why Kerr (1937, p. 545) obtained a parallel orientation of clay in both fresh and salt water.

Chang (1941, p. 220–221) suggested that, because flocculation resulted in a random arrangement of the particles, dried electrolyte-flocculated clays should show a random orientation. However, he neglected to consider drying or compaction forces.

Effect of organic matter.—The writer's experiments with dispersed clay and organic matter obtained from black shales show that either gravity settling or flocculation gives a parallel orientation of the clay particles. Analyses for organic content also show that fissility and the presence of organic matter are closely related. The mechanism for this is not clear. Perhaps the organic particles are flaky, or perhaps the organic matter is in solution or is a fine colloid so that the adsorption of the organic matter does not prevent or disturb the parallel arrangement of the clay particles. Sideri (1936, p. 471; 1938, p. 131) showed that the moistening of a clay with a humate solution resulted in the formation of a number of parallel laminae.

In some situations organic matter may hinder the development of a platy structure (Peterson, 1944, p. 44). However, of the mudrocks studied, none of the massive ones had even a moderate amount of organic matter. Peterson worked with puddled clays. As his clays were not dispersed, perhaps there was no opportunity for the humus to be adsorbed, and the excess combined as a cement around clay-humus aggregates.

The flaggy shales contain by far the largest amount of organic matter. Humus is almost irreversibly adsorbed by clay. The resulting cementing action enables the shale to break into large sheets without falling apart. Campbell's observation (1946, p. 832), that in the New Albany shale, laminae are thinnest where the amount of carbon is the greatest, substantiates this view.

The black flaggy shales show a lineation of flattened white clay or clay-calcite lenses in an opaque organic matrix. Why these clay lenses are unstained is not apparent.

Effect of iron.—Gravity settling of a red-clay suspension, flocculation of clay and $Fe(OH)_3$ together, and flocculation of clay with powdered Fe_2O_3 all produce a random orientation of the clay particles. Apparently the $Fe(OH)_3$ and the Fe_2O_3 enmesh or adsorb the clay particles on irregular surfaces hindering their normal orientation.

A parallel orientation was obtained in a red mudrock by flocculating a clay suspension with

a $FeCl_3$ solution. The ferric ions could have been adsorbed uniformly by the clay and later oxidized forming small grains of Fe_2O_3 which would not disturb the orientation. However, other explanations for red fissile shales are not excluded. If only a small amount of $Fe(OH)_3$ were flocculated with the clay, orientation might not be disturbed. It is also possible that sufficient iron to give a red color upon oxidation might diffuse into a fissile shale.

Other workers have observed the disturbing influence of iron on the structure of mudrocks. Keller (1936, p. 55) saw only a slight orientation in a flocculated lateritic soil. Sideri (1938, p. 132–134) reported that "anisotropic colloidal particles of kaolinite lose their ability to distribute themselves in parallel rows" when sesquioxides are present. After the removal of iron oxides with Tamm's Reagent, the formation of parallel structure began. Peterson's experiments (1944, p. 45) with puddled kaolinite showed a gradual decrease in the platiness of the dried residue as the amount of Fe_2O_3 increased.

Effect of aluminum.—The influence of aluminum on the structure of mudrocks is similar to that of iron; the only difference is that a flocculated $Al(OH)_3$-clay system gave a parallel orientation of the clay. The alumina gel did not trap the clay so as to prevent its orientation upon drying. Flocculation of an Al_2O_3-clay system gave a random structure similar to the Fe_2O_3-clay system.

Effect of silica.—The drying of a silica gel in which clay had been dispersed produced a hard, chertlike mass in which much of the clay had been expelled to the bottom where the orientation was parallel to the bottom. Apparently silica gel exerts a force which tends to clear it of impurities as it dries. Powdered amorphous SiO_2 in a clay suspension produces the same random structure as Fe_2O_3 and Al_2O_3.

Post-Depositional Changes

Compaction

Flocculated clays can hold up to 85–90 per cent water by volume. Hedberg's detailed work (1936, p. 269) on the compaction of clay showed that a mechanical rearrangement of the clay particles to a parallel position takes place early during compaction and is usually complete in the top 5–10 cm. of a mud deposit. He made no mention of the influence of impurities in the clay.

Penecontemporaneous Clay Minerals

Keller (1946, p. 70), Payne (1942, p. 1766), and others suggested that clay minerals may form in a mud during diagenesis. Keller used the mechanism of random growth of clay minerals in a gel to explain massive fireclays in Missouri. Allen's observations (1945, p. 265–266) on the migration of clay in colloidal suspension or in solution increase the possibility of such a mechanism.

Payne (1942, p. 1717) and Grim and Cuthbert (1945, p. 94) inferred that the diagenetic formation of illite in marine sediments explained the fissility of most shales. As the writer found no relationship between fissility and the presence of illite, this view cannot be substantiated.

Wave Action

Weller (1930, p. 729) suggested that repeated stirring of bottom muds by wave action may explain some thick-bedded, structureless mudstones. In the writer's flocculation experiments, repeated stirrings of the flocculated clays produced no effect on the final structure. But if consolidated or partially consolidated shales were broken by wave action and reassembled so as to break up the common orientation of the clay particles, a mudstone might be produced. The probability of this happening appears to be small though such a fabric was observed in one of the samples studied.

Weathering

The weathering of mudrocks determines to a large degree the way a potential fissility will be expressed. In general, weathering of a given shale increases the platiness of fracture fragments. This relation was noted in one sample that was flaggy when weathered and more massive when unweathered. McKelvey (1946, p. 28–29) observed the same phenomena in the Phosphoria formation of Wyoming and Utah. Soft, black fissile shales at the surface became hard, massive siltstones at a depth of

30 feet at Sublette Ridge, Utah. Investigations in mines of this area showed all rocks to be harder and more massive with depth. Mudrocks with a parallel arrangement of the clay particles have a potential fissility which may not be realized until weathering has weakened the cementing materials.

It is also entirely possible that extreme weathering may destroy fissility by changing a shale into a soft clay. The differential swelling and shrinking of a clay caused by alternate wetting and drying can disrupt the orientation pattern. Mohr (1932, p. 358–359) described this condition in the weathering of marine shales in the Netherlands East Indies. This may explain Bain's statement (1896, p. 293) that "the absence of shaly parting is often a matter of weathering as shown by the greater predominance on fresh exposures."

Compaction and later elastic recovery caused by erosional unloading undoubtedly aids the development of fissility.

Conclusions

(1) The breaking characteristics of mudrocks can be described as massive, flaky, flaggy, or some combination of these three pure types.

(2) Fissility is associated with a parallel arrangement of the clay particles; massiveness, with a random arrangement.

(3) Organic matter increases the tendency toward parallel arrangement of the clay particles.

(4) Colloidal sesquioxide and silica particles increase the tendency toward a random arrangement of the clay particles. Ferruginous fissile shales can be explained by the oxidation of iron ions entrapped during settling or flocculation or introduced epigenetically by solutions.

(5) Keller's explanation (1946, p. 70) of some mudstones as originating by the random growth of clay minerals in a gel is a probable one.

(6) The nature of the cementing agent determines whether a shale will be flaky or flaggy. If the cementing agent can hold the material together in a large slab, the shale will be flaggy. If the amount or the tenacity of the cementing agent is small, the shale will be flaky. As colloidal organic matter is almost irreversibly adsorbed and dehydrated, black shales predominate among flaggy shales; other well-cemented shales may be somewhat flaggy.

(7) Other factors being equal, most cementing agents of mudrocks cause a decrease in fissility.

(8) The influence of composition on the type of fissility is reflected in the color. Flaggy shales are predominantly black or gray black; flaky shales, gray or grayblack; and mudstones, white, gray, red, yellow, purple, etc.

(9) Except where the clay has obtained a random orientation by adsorption on particles of sesquioxides or silica, compaction of muds results in a rotation of the particles to a position perpendicular to the force of compaction.

(10) Moderate weathering increases the fissility of a shale by the partial removal of the cementing agents along planes parallel to the bedding or by expansion caused by hydration of the clay particles. Intense weathering produces a soft massive clay.

(11) Given a certain critical, but small, amount of clay, the type of fissility is independent of the particle sizes in a mudrock.

(12) With a random collection of mudrocks, the type of fissility is uncorrelatable with the kind of clay minerals present. Possibly, however, under certain conditions, the kind of clay mineral may determine the type of fissility.

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CHLORINE IN IGNEOUS ROCKS

SOME ASPECTS OF THE GEOCHEMISTRY OF CHLORINE

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ABSTRACT

The old value of 0.048 per cent chlorine for the average in the earth's crust is considered too high, and an average of 0.02 per cent appears more probable. The average chlorine content of approximately 100 granitic rocks from North America and other parts of the world was found to be 0.022 per cent; of 20 gabbros, 0.021 per cent; and of 40 basalts and diabases, 0.014 per cent. Chlorine in ultramafic rocks may range from 0.005 to 0.2 per cent or more; alteration (as by autometamorphic processes) may account for the high chlorine contents.

A large fraction of the total chlorine in common igneous rocks occurs in hydroxyl minerals such as biotite and hornblende; presumably chloride ion replaces hydroxyl ion. These minerals may contain several tenths of a per cent of chlorine. In basic rocks containing much phosphorus (>1 per cent P_2O_5), apatite may account for much or most of the chlorine, but otherwise there is no close relation between phosphorus and chlorine. Some chlorine is also present as chloride in liquid (aqueous) inclusions, and the chlorine in such anhydrous minerals as the feldspars is probably present largely in this form.

Chlorine does not show any marked trend in the differentiation of the common igneous rocks. Since it is an element of well-developed atmophilic and hydrophilic characteristics, a general lack of correlation between chlorine and the major and minor constituents of rocks is not unexpected; an appreciable, perhaps large, fraction of the chlorine in a magma escapes on crystallization.

The ratio Cl/H_2O+ shows a general increase from subsilicic rocks (basalts and gabbros) to silicic. This trend may be attributed to enrichment of Cl compared to H_2O in the residual liquid as differentiation proceeds to more silicic products.

The general geochemistry of chlorine is discussed. Few elements have siderophilic, lithophilic, hydrophilic, and atmophilic properties all developed in comparable degree.

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Introduction

Although a rather large number of values for the chlorine content of igneous rocks is scattered through the literature, many of them pertain to uncommon rock types. Data for granites, gabbros, and basalts, which geochemically are of greater importance, are inadequate. For such rocks, "tr" and "none" have frequently been reported, and it is difficult to arrive at an average value. Improved methods for the determination of low concentrations of chlorine now make it easily possible to obtain the chlorine contents of such rocks.

The aim in the present work has been to obtain the chlorine content of the more common rock types by the use of individual and composite samples. Some study has been made of the distribution of chlorine among the component minerals of igneous rocks.

Except for two gravimetric chlorine values obtained from the Minnesota Rock Laboratory, all the new results reported here are from analytical work by Kuroda.

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Analytical Method

Most of the chlorine determinations were made by a photometric method in which the sample is fused with sodium carbonate, chloride is precipitated as silver chloride, the washed precipitate is dissolved in ammonia, and the resulting solution is treated with sodium sulfide to give a brown silver sulfide sol whose optical density is measured (Kuroda and Sandell, 1950). A few determinations were made by a potentiometric method (Kolthoff and Kuroda, 1951). The error in the photometric method usually does not exceed 20 per cent for 0.01–0.02 per cent Cl and 10 per cent for 0.05 per cent Cl.

Care was taken to avoid possible chlorine contamination in the laboratory. Filter papers were washed with water before use, and blanks were run through the whole procedure. Conductivity water was used in all determinations. Except in certain drill core samples, and possibly one or two other samples, foreign chloride probably was not present in significant amounts in the rocks analyzed.

It is believed that no halogen was introduced by the bromoform separation of minerals. The fractions were thoroughly washed with pure ethyl alcohol. A biotite separated from a granite by hand picking gave the same chlorine content as a sample obtained by bromoform separation.

Previous Data

Meteorites

Results of determinations of chlorine in meteorites are summarized in Table 1. It is difficult to derive reliable averages from these data. For silicate meteorites an average content of some hundredths of a per cent chlorine appears probable on the basis of the various values reported. Goldschmidt (1937, p. 64) accepted Merrill's 1916 average of 0.08 per cent for silicate meteorites, but for some reason did not mention Merrill's 1930 average of 0.03 per cent. Although the average chlorine content of meteorites has not been well established, it appears that the silicate phase contains chlorine in amounts comparable to that in igneous rocks.

The chlorine content of iron meteorites is even more uncertain. Chlorine may occur in relatively large amounts as lawrencite ($FeCl_2$) in iron meteorites (*e.g.*, 1.5 per cent Cl in the Lime Creek meteorite). Merrill's (1930, p. 9) average of 0.02 per cent Cl is based on very few analyses and, because of the sporadic distribution of chlorine, cannot be given much weight. Goldschmidt (1937, p. 100) takes 0.1–0.15 per cent as the possible average chlorine content of meteorites as a whole (10 parts silicate, 1 part troilite, 2 parts nickel-iron), thus assigning by inference as much as 0.5 per cent Cl to iron meteorites. The chlorine content of the sulfide phase is unknown, but presumably is small.

Igneous Rocks

The value usually quoted for the average chlorine content of the lithosphere is that of Clarke and Washington (1924, p. 16), namely 0.048 per cent. Although the average is reported to two significant figures, it is evident from the method of averaging that the mean must be very approximate. These authors first divided the sum of the percentages of chlorine by the number of samples in which chlorine had been determined (857) and obtained 0.096 per cent. This average would be much too high because of the preponderance of unusual rocks higher in chlorine than the average. But, if the sum of percentages of chlorine from 857 samples is divided by 5159, the number of rocks in which the major constituents had been determined, the obviously low value of 0.0016 per cent Cl is obtained. Therefore, these authors took the mean of the high and low values, thus obtaining 0.048 per cent Cl for their final average.

Selivanov (1940, p. 809) found an average of 0.025 per cent Cl in 4 Russian gabbros and basalts, 0.028 in 4 granites, 0.054 in 2 granodiorites, and a grand average of 0.031 per cent in 23 rocks, including those mentioned and other less common rocks.

Selivanov's average for igneous rocks as a whole is accepted by Rankama and Sahama (1950, p. 756) in preference to Clarke and Washington's figure.

Tröger (1934, p. 302–307), making use of earlier data, has calculated chlorine (and fluorine) averages for different rock types. He reports a mean of 0.01 per cent Cl for rocks of the granite-quartz diorite group, 0.03–0.05 per cent for the diorite-gabbro group, and 0.01–0.02 per cent for the syenite-monzonite group.

TABLE 1.—CHLORINE IN METEORITES
In per cent

Type	Cl
Silicate meteorites	0.08*, 0.03†
Silicate phase of silicate meteorites	0.03†, 0.09‡
Chondrites	0.047§
Iron meteorites	0.02¶
Metal phase of silicate meteorites	tr¶

* Merrill (1916, p. 28); average of 5.

† Based on superior analyses (Merrill, 1930, p. 17). The metallic and troilite phases amount to about 15 per cent.

‡ I. and W. Noddack (1930, p. 757); based on 42 stony meteorites.

§ I. and W. Noddack (1934, p. 178).

¶ Average of superior analyses (Merrill, 1930, p. 9).

Many chlorine determinations are recorded in the publications of the United States Geological Survey and are conveniently available in the collections of analyses by Clarke (1915) and Wells (1937), covering the period 1880–1936. From these data it is especially evident that chlorine tends to be enriched in alkalic rock types.

Sedimentary Rocks

Very little is known about the chlorine content of sediments. Stokes (*in* Clarke, 1924, p. 30) found 0.02 per cent Cl in a composite of 345 limestones and a trace in a composite of 253 sandstones. Determinations of chlorine in metamorphic rocks are also very scarce.

CHLORINE IN INDIVIDUAL SAMPLES OF COMMON ROCK TYPES

In order to learn the approximate range of chlorine in common types of rocks, a number of individual samples from various localities were analyzed. The results for granitic rocks,

gabbros, andesites, diabases, and basalts are plotted in a frequency diagram (Fig. 1). Some of these values are from analyses of rock series are distinguished in the diagram. In addition to granites and granite gneisses, this group includes granodiorites, quartz monzonites,

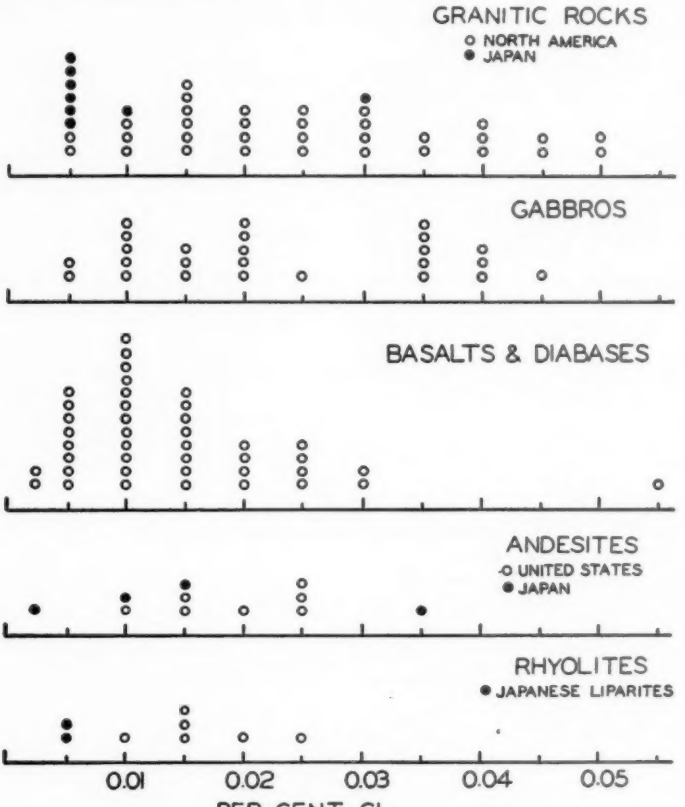

FIGURE 1.—FREQUENCY DIAGRAM OF CHLORINE IN COMMON IGNEOUS ROCKS
A Minnesota diabase with 0.16% chlorine is not shown

(Tables 7, 8, 9, 10, 11) made for the purpose of tracing the trend of chlorine in igneous differentiation.

GRANITIC ROCKS: The samples represented in Figure 1 are distributed in origin as follows: 16 from Minnesota (mostly from the east-central part), 4 from Missouri, 7 from Texas (Llano region), 4 from unspecified North American localities, 1 from China, and 8 from Japan. The American samples are mostly, if not all, of Precambrian age. Most of the Japanese specimens are markedly lower in chlorine than the U. S. samples, and the two regions and one hornblende tonalite (0.045 per cent Cl).

GABBROS: The localities of the samples are as follows: Minnesota, 5; Michigan, 3; New York, 2 or more; Colorado; 1, Wyoming, 1; Montana, 1; Africa, 1; Antarctica, 1; and 8 from unspecified North American localities.

ANDESITES: Seven samples are from the U. S. and 4 from Japan.

BASALTS AND DIABASES: Twelve samples are from Minnesota, 4 from Michigan (Greenstone flow samples counted as one sample), 6 from West Texas, and the remainder of the 40 samples from various U. S. localities, with

the exception of 1 Deccan and 1 Antarctic basalt. The following individual chlorine values may be of interest: diabase, Triassic, 0.014; basalt, Columbia River, 0.014; trap, Deccan, 0.004; basalt, Antarctica, 0.009; basalt dike in St. Cloud, Minn., granodiorite, 0.016; pillow basalt, 0.003; greenstone, Winton, Minn., 0.013 per cent.

TABLE 2.—CHLORINE IN ULTRAMAFIC ROCKS
In per cent

Rock	Cl	H_2O+
Dunite, Balsam, N. C.	0.006	—
Peridotite	0.006	0.06
Dunite, unspecified U. S. locality	0.012	8.7
Dunite, N. C.	0.014	0.78
Kimberlite, Africa	0.014	6.50
Pyroxenite, Lewis Hills, Newfoundland	0.020	0.51
Bronzitite, Beartooth Mtns., Montana	0.054	0.46
Harzburgite, Beartooth Mtns., Montana	0.054	6.17
Harzburgite, Lewis Hills, Newfoundland	0.06	10.05
Dunite, feldspathic, Lewis Hills, Newfoundland	0.10	11.6
Harzburgite, serpentinized, unspecified North American locality	0.13	8.43
Dunite, serpentinized, Lewis Hills, Newfoundland	0.24	14.08
Serpentine, Akita Prefecture, Japan	0.002	
Serpentine, Gifu Prefecture, Japan	0.002	
Serpentine, Saitama Prefecture, Japan	0.003	
Serpentine, Hoboken, N. J.	0.039	
Serpentine, San Francisco, Calif.	0.041	
Serpentine, Yearlying Head, Llano Co., Texas	0.050	
Serpentine, U. S. or North America	0.052	

ULTRAMAFIC ROCKS: The chlorine content of samples of this group (including serpentines) is listed in Table 2. The H_2O+ content, where known, is also listed as having a bearing on the interpretation of the results.

SEDIMENTARY AND METAMORPHIC ROCKS: A detailed study of these rocks lay outside the scope of the present work, but a few specimens were analyzed to obtain an estimate of the probable chlorine content (Table 3).

TABLE 3.—CHLORINE IN SOME SEDIMENTARY AND METAMORPHIC ROCKS
In per cent

Rock	Cl
Shale (Cambrian), St. Croix Falls, Wis.	0.014
Residual clay, Redwood Falls, Minn.	0.007
Slate (Thomson formation), Carlton, Minn.	0.007
Slate (Rove), Gunflint Lake, Minn.	0.023
Phyllite, Morrison Co., Minn.	0.008
Staurolite schist, Morrison Co., Minn.	0.007
Mica schist, Southern Aitkin Co., Minn.	0.008
Graphite gneiss, Gifu Prefecture, Japan	0.014
Hornblende gneiss, Gifu Prefecture, Japan	0.023
Sandstone (St. Peter), Minneapolis, Minn.	0.002

CHLORINE IN COMPOSITES OF IGNEOUS ROCKS

Determinations of chlorine in individual samples were supplemented by determinations in composite samples prepared on the basis of rock class, silica percentage, or region. Equal weights of individual samples were taken in preparing the composites. The results are given in Table 4. A few of the averages were obtained from the results of individual determinations, as in the case of the central Texas granites and the Greenstone flow of Michigan. The water contents and the weight ratios Cl/H_2O+ are also given for later consideration.

AVERAGE CHLORINE IN VARIOUS ROCK TYPES

GRANITIC ROCKS: The grand average, based on 99 samples (individual and composite analyses) weighted equally, is 0.022 per cent Cl. This may be compared with the average of 0.024 per cent in 16 Minnesota granitic rocks.

GABBROS: The average of 20 samples (Nos. 3161, 3163 and 3166 of Table 8 counted as one sample) is 0.021 per cent Cl.

BASALTS AND DIABASES: The average of 39 samples is 0.014 per cent Cl. The average of 12 Minnesota samples is 0.011 per cent. The

Minnesota diabase containing 0.16 per cent Cl (duplicate analyses) has not been included in this average. If it is included the average becomes 0.018 per cent.

be seen that chlorine rarely reaches 0.05 per cent in such rocks.

The average value of 0.02 − per cent adopted here must be considered approximate and

TABLE 4.—CHLORINE IN COMPOSITES OF IGNEOUS ROCKS
In per cent

Silica range	Average SiO_2	Rock type and number of samples	Region	Cl	H_2O+	Cl/H_2O+
80.6–76.3	77.7	Granite (12)	U.S.A.	0.024	0.22	0.11
81.5–75.1	77.0	Granite (8)	U.S.A. and Canada	0.024	0.23	0.10
77.4–71.5	74.1	Granite, etc. (9)	California	0.038	0.23	0.17
76.8–68.2	73.1	Granite (7)	Texas (Llano region)	0.017	0.23	0.08
74.3–70.6	72.6	Granite (7)	Canada	0.016	0.42	0.04
73.9–70.0	72.0	Granite (9)	Africa	0.017	0.57	0.03
69.6–65.5	68.4	Granitic (14)	Various	0.018	0.60	0.03
62.9–62.1	62.6	Diorite (3)	California	0.045	0.60	0.075
65.7–52.7	61.5	Diorite (4)	Oregon	0.035	1.41	0.025
	47.6	Gabbro-diorite (4)	California	0.028	1.24	0.023
50.4–46.0	47.4	Basalt	Michigan (Greenstone flow)	0.013	2.2	0.006
64.1–56.0	60.3	Syenite (6)	N.Y. and Minn.	0.055	0.66	0.08
57.9–49.6	53.1	Phonolite (4)	Various	0.050	2.6	0.02

ULTRAMAFIC ROCKS: The average of 13 samples of this type (Table 2, serpentines not counted) is 0.06 per cent. Because of the small number of samples and the variability in chlorine content the exact figure is of little significance.

OTHER ROCKS: The following averages were based on a small number of samples and are to be looked upon as very approximate: rhyolites (8 samples), 0.014; andesites (11), 0.018; diorites (7), 0.04; syenites (6), 0.05+; phonolites (4), 0.05 per cent Cl.

AVERAGE CHLORINE CONTENT OF THE LITHOSPHERE

Thus we may take the average chlorine content of granitic rocks to be 0.02+ per cent and that of basaltic rocks to be 0.015 per cent, or say, 0.02 − per cent. Since these rocks are of predominant importance in the outer layer of the earth, the value 0.02 − per cent may be assigned as the average for chlorine in the upper lithosphere. The value 0.048 per cent given by Clarke and Washington (1924, p. 16) is doubtless too high, because of insufficient weighting of the common rocks. From Figure 1 it will

tentative. A larger number of samples, from all parts of the world, must be examined to obtain a more exact average. More analyses of basaltic rocks especially are needed. Almost half the basalts and diabases represented in the average for these rocks are from Minnesota and Michigan, and they may unduly affect the mean and give an appearance of greater uniformity than acually exists.

REGIONAL FACTOR IN OCCURRENCE OF CHLORINE

There is some evidence that the chlorine content of igenous rocks of certain regions may differ appreciably from the average of the class. Some of the Japanese rocks examined in the present work have an abnormally low chlorine content. Thus six of the eight Japanese granites analyzed contain only 0.003 to 0.005 per cent chlorine (Table 5), or much less on the whole than granitic rocks from the United States, Canada, and Africa (Table 4; Fig. 1). The Japanese serpentines analyzed (Table 2) are also very low in chlorine (0.002 per cent or less), in contrast to serpentines from various parts of the United States (0.05 per cent).

Most of the other Japanese rocks analyzed have normal chlorine contents, except possibly the liparites. Chemical analyses of the Japanese rocks are not available, so it is not known whether the chlorine abnormality can be correlated with any other peculiarity.

On the other hand, some areas of granitic rocks are richer in chlorine than the average as found in this work. Thus the potassic rapakivi granites of Finland contain, on the average, 0.06 per cent chlorine, as well as 0.36 per cent fluorine (Sahama, 1945a, p. 44). The high chlorine content of these granites is perhaps connected with high potassium, so that in a strict sense it perhaps should not be ascribed to the effect of a regional factor. The Russian and Siberian igneous rocks examined by Selivanov (1940, p. 809) show normal amounts of chlorine.

Behavior of Chlorine in Differentiation of Rocks

Chlorine shows no marked preference for silicic or subsilicic rocks. There is some indication that its amount in basaltic rocks is less than in other rocks. Figure 1 shows there is a tendency for the chlorine contents of basalts and diabases to cluster around 0.01 per cent, and the range is, on the whole, more restricted than in the plutonic rocks, whether acidic or basic. If the Japanese granites are disregarded, there is a distinct difference in the distribution pattern of chlorine content in basalts and diabases, on the one hand, and granitic rocks on the other. It is perhaps not surprising that the scatter should be greater in plutonic rocks where conditions for differentiation are more favorable. There is a suggestion in Figure 1 that extrusive rocks may be lower than plutonic rocks in chlorine. Many ultramafic rocks are quite high in chlorine, much more so than any other rock type examined in this work, although the variability is pronounced (from 0.006 to 0.24 per cent Cl in 13 samples). There is some indication that the chlorine content tends to rise with the water content (Table 2), but the relationship is not close. Serpentines from four United States localities contained 0.04–0.05 per cent chlorine, but three Japanese specimens contained only a few thousandths of a per cent. The reason for the frequently high chlorine content of ultramafic rocks is not certain, but possibly the element has been introduced by the action of aqueous chloride solutions after primary crystallization, and is present in serpentinization products. This solution might well be residual liquid from the crystallization of the magma. Sahama (1945b, p. 58) has explained the presence of appreciable amounts of boron in Finnish serpentine rocks by autometamorphism following an enrichment of boron in residual aqueous solutions.

Table 5.—Chlorine in Japanese Igneous Rocks*

In per cent

Rock	Locality (Prefecture)	Cl
Liparite	Ishikawa	0.005
Liparite	Yamagata	0.005
Pitchstone	Nagano	0.045
Perlite	Nagano	0.10
Granite	Yamanashi	0.003
Granite	Gifu	0.005
Biotite granite	Okayama	0.005
Biotite granite	Gifu	0.003
Biotite granite	Nagano	0.030
Biotite granite	Ibaragi	0.003
Hornblende granite	Gifu	0.011
Hornblende granite	Fukushima	0.003
Pyroxene andesite	Nagano	0.002
Pyroxene andesite	Gifu	0.037
Augite andesite	Nagano	0.013
Augite andesite	Kumamoto	0.010
Diabase	Gifu	0.011
Diorite	Yamanashi	0.018
Gabbro-diorite	Gifu	0.010
Gabbro-diorite	Saitama	0.015

* For chlorine in Japanese serpentines see Table 2.

Glasses as a rule are high in chlorine as well as fluorine. Shepherd (1940, p. 119–120) has reported the contents of these elements in glasses from California and New Mexico. In 11 samples, in which both halogens were determined, the average is 0.08 per cent Cl (range 0.03 to 0.13) and 0.07 per cent F (range 0.03 to 0.12). Bowen (1937, p. 14) has reported even higher contents in glasses of the East African alkalic lavas; the maximum chlorine contents of 0.40 and 0.41 per cent were found in two rhyolitic obsidians; the respective

fluorine percentages are 0.17 and 0.68. The H_2O+ content of these samples is only about 0.1 per cent; high chloride does not require a high water content. In the present work, six glasses were analyzed for chlorine: obsidian (black), Yellowstone Park, 0.12 per cent; rhyolitic obsidian, California, 0.031 per cent; two volcanic glasses, west Texas, 0.051 and 0.055; and pitchstone and perlite, Nagano Prefecture, Japan, 0.045 and 0.10 per cent, respectively. Five of the six samples have a higher chlorine content than the average of their equivalent crystallized rocks.

In glasses, chlorine is doubtless present as a metal chloride in solution, and the original melt is probably closely represented. If such a melt had crystallized, comparatively little of the chlorine might have entered the crystals, and the remainder, as a metal chloride or hydrogen chloride, could have escaped or have been carried off by water emanating from the magma, or perhaps by later solutions. It seems likely that silicic, and possibly other, magmas were considerably richer in chlorine than the rocks now representing them.

There is some indication that in glassy rocks chlorine occurs in amounts not greatly different from fluorine, whereas in crystallized rocks chlorine tends to occur in smaller amounts, as if more chlorine than fluorine has escaped during crystallization.

As is well known, the chlorine content of alkalic rocks is frequently high, running up into tenths of a per cent or even approaching 1 per cent. When the concentrations of sodium and chlorine are both high, sodalite may be formed. Evidently chlorine tends to be concentrated with the alkalies, particularly sodium; it may also be that chlorine is less readily lost from such magmas and most of it finds its way into the rock.

There is a difference in the behavior of

TABLE 6.—CHLORINE IN ROCKS OF GREENSTONE FLOW, MICHIGAN

In per cent

Broderick's No.	Cl	Cl/H_2O+
3132 (top of flow)	0.008	0.004
3134	0.010	0.005
3136	0.018	0.010
3138	0.014	0.008
3140	0.016	0.012
3142	0.015	0.007
3144	0.009	0.005
3146	0.019	0.008
3148	0.020	0.008
3150	0.008	0.004
3152	0.009	0.003
3154 (base of flow)	0.010	0.004
Average	0.013	0.006

TABLE 7.—CHLORINE IN ROCKS OF ENDION SILL, DULUTH

In per cent

	Cl	Cl/H_2O+
Ophitic diabase (No. 1)	0.019	0.012
Diabase (gradation to red rock) (No. 3)	0.010	0.005
Intermediate red rock (No. 4)	0.011	0.009
Red rock (No. 6)	0.011	0.006
Rhyolite (above sill) (No. 7)	0.011	0.046

TABLE 8.—CHLORINE IN MISCELLANEOUS MICHIGAN KEWEENAWAN ROCKS

In per cent

Broderick's No.	Rock*	Cl	Cl/H_2O+
3091	Melaphyre	0.042	0.018
3095	Ophite	0.003	0.001
3103	Glomeroporphyrite	0.018	0.005
3104	Melaphyre	0.016	0.007
3106	Melaphyre	0.019	0.006
3110	Glomeroporphyrite	0.049	0.016
3127	Melaphyre	0.020	0.010
3129	Melaphyre	0.010	0.005
3156	Felsite (intrusive)	0.009	0.027
3157	Felsite	0.010	0.043
3159	Felsite	0.005	0.020
3161	Gabbro aplite	0.026	0.018
3163	Gabbro, intermediate	0.034	0.015
3166	Gabbro, basic	0.040	0.018
3169	Felsite porphyrite	0.024	0.011
3170	Porphyrite	0.012	0.005

*Nos. 3091, 3095, 3103, 3104, 3106, 3110, 3127, and 3129 are drill-core samples, and the possibility of foreign chloride in these is not excluded. The other samples are free from this uncertainty.

chlorine and water in the differentiation of the calc-alkaline rock series. The ratio Cl/H_2O+ generally increases with increasing SiO_2 content. For basalt and diabase this ratio is approximately $0.014/1.4 = 0.01$[1]. For gabbro it appears to be higher, approximately $0.021/1.1 = 0.02$[2]. It rises to 0.03–0.04 for the common granites and allied rocks and to ~0.1 for the high-silica granites (Table 4). The increase in the ratio is due largely to the lower water content of the more silicic rocks, but in part to a slight increase in chlorine in the same direction. A tentative explanation is offered for the change in the ratio. Chloride ion enters the common hydroxyl minerals of igneous rocks, presumably by replacing OH^- in the structure. The empirical ionic radius of Cl^- (1.81A) is so much larger than that of OH^- (1.33A) that it seems likely Cl^- will be enriched with respect to water in the liquid on progressive crystallization. The increasing value of the ratio Cl/H_2O+ in the liquid results in an increase of the ratio in later fractions of the hydroxyl-containing minerals. In the simplest case, the chlorine content of a rock, as a function of OH^- or H_2O+, would depend upon the water content of the rock and the stage of the fractionation. Actually there are many other variables involved: variation of the fractionation factors, entry of chlorine into minerals by means other than isomorphous substitution, escape of water and chlorine from the magma, the original composition of the magma, etc. No close relation can be expected between the water and chlorine contents of an igneous rock.

TABLE 9.—CHLORINE IN VOLCANIC ROCKS OF CLEAR LAKE AREA, CALIFORNIA*

In per cent

	Cl
Olivine basalt	0.054
Augite andesite	0.018
Rhyolitic obsidian	0.031
Pyroxene rhyodacite	0.051
Silicic pyroxene dacite	0.019
Biotite-hornblende rhyodacite	0.033
Biotite rhyolitic pumice	0.043
Olivine dacite	0.015
Hypersthene dacite	0.014
Olivine andesite	0.025

* These rocks have been described by Anderson (1936).

TABLE 10.—CHLORINE IN WEST TEXAS ROCKS

In per cent

	Cl	P_2O_5	Cl/P_2O_5
Sheep Canyon basalt	0.030	3.0	0.010
Sheep Canyon basalt	0.012	0.87	0.014
Elephant Mountain basalt	0.015	0.96	0.015
Cottonwood Springs basalt	0.021	1.97	0.011
Cottonwood Springs basalt	0.020	1.81	0.011
Butcherknife Hill basalt	0.011	1.10	0.010
Buck Hill syenite	0.021	0.28	0.075
Potato Hill andesite	0.026	1.13	0.023
Crossen trachyte	0.014	0.07	0.2
Mitchell Mesa rhyolite tuff flow	0.024	0.03	0.8
Pruett tuff	0.015	0.04	0.4
Duff tuff	0.010	0.08	0.13
Vitrophyre, Tascotal Mesa (F = 0.34%)	0.20	0.00	—

CHLORINE IN ROCK SERIES: Since chlorine shows no marked preference for rocks high or low in silica, it is perhaps not surprising that no well-defined trend can be traced in closely related rocks of the series in Tables 6–11.

No correspondence between chlorine content and the major or minor constituents of the rocks of the Greenstone basalt flow of northern Michigan (Table 6) could be found. The only relationship noted, which may or may not be significant, is the generally lower chlorine content of the two uppermost samples and the three lowest samples as compared to the seven samples closer to the center of the flow. There is no tendency for chlorine to be concentrated in the pegmatitic doleritic horizons (Nos.

[1] The average H_2O+ content of the basalts and diabases in which Cl was determined is not known. Therefore, 1.8, the average total water percentage of plateau basalt as given by Daly (1933, p. 17) has been somewhat arbitrarily multiplied by ¾ to obtain the percentage of H_2O+. The Greenstone flow of Michigan (Table 6) and the basic members of the miscellaneous Keweenawan (Table 8), for which H_2O+ is known, have Cl/H_2O+ equal to 0.007 and 0.01.

[2] The total water content of gabbro is taken to be 1.45 per cent, and H_2O+ as ¾ × 1.45 = 1.1 per cent.

3136 and 3140), in which, as found by Broderick (1935, p. 512), SiO_2, Fe, Na, K, Ti, P, S, Mn, and Cu are higher than the average, and Al, Mg, and Ca are lower.

TABLE 11.—CHLORINE AND FLUORINE IN GRANITIC ROCKS OF LLANO REGION, TEXAS

In per cent

	Cl	F*	F/Cl
Quartz monzonite, Town Mtn.	0.011		
Granite, Petrick	0.030	0.12	4
Granite, Granite Mtn.	0.019	0.14	7
Granite, Cassaday	0.014		
Granite porphyry (llanite), Babyhead	0.017	0.25	15
Granite, Bear Mtn.	0.014	0.06	4
Alpite, Petrick	0.015	0.13	9

* Fluorine values taken from Goldich (1941, p. 700). Fluorine was determined only in those rocks in which fluorite is present.

Although these samples are from drill cores, there does not appear to be appreciable contaminating chloride present. This conclusion is based on the small proportion of water-soluble chlorine present. Seven of the 12 samples extracted with water showed 0.000 to 0.003 per cent Cl (based on the rock) in the filtrates. Such amounts of chlorine can be extracted from similar rocks not subjected to possible contamination. On the other hand, samples of the Kearsarge flow of Michigan (Broderick, 1935, p. 521), also drill cores, showed uniform total chlorine contents of 0.04–0.05 per cent, of which about three-fourths was water soluble. This suggests chlorine contamination, and we therefore omit the individual values for this series.

In the Endion sill of Duluth, Minnesota, studied by Schwartz and Sandberg (1940, p. 1139), the lower diabase is highest in chlorine (Table 7). There is no increase in chlorine in the upper "red rock" acidic facies; on the contrary there is a significant decrease. If chlorine partitioned between early crystals and liquid in comparable amounts, it was not incorporated in the later upper rocks and roughly half of the total amount was lost, presumably by escape of aqueous solutions. Although no enrichment of chlorine is shown in the red-rock facies of the Endion sill, the opposite seems to be true of the Duluth gabbro. Two samples of more or less typical gabbro showed 0.012 and 0.022 per cent Cl, whereas a sample of orthoclase gabbro (Duluth) showed 0.036 per cent Cl and an augite syenite (Cook County, Minn.) 0.041 per cent.

Table 8 shows that the chlorine contents of the Mt. Bohemia intrusive (Nos. 3161, 3163, 3166) of Michigan decrease from basic to intermediate to aplite phases. The felsite of Bare Hill (3156), regarded as intrusive, contains 0.009 per cent chlorine, or rather less than the average of the basaltic rocks of the region. Samples of the lower (3157) and upper (3159) halves of the Mt. Houghton felsite contain 0.010 and 0.005 per cent Cl. Again the relatively low chlorine contents of these acidic (74.51 and 74.23 per cent SiO_2) rocks are to be noted. The chemical analyses of the two samples are much the same, but there seems to be significantly less chlorine in the rock of the upper part of the mass.

In the west Texas basalts (Table 10) studied by Goldich and Elms (1949, p. 1143), a correspondence between chlorine and phosphorus is evident; Cl/P_2O_5 ranges from 0.01 to 0.015 in six samples. These rocks are unusually high in phosphorus. Even if it were not known that chlorine can occur in apatite, it could be concluded that there is a mutual relation between phosphorus and chlorine. In the andesite the ratio Cl/P_2O_5 is 0.023 or about twice that in the basalts. In the other rocks, the phosphorus percentage is low (highest in the syenite, 0.28 per cent P_2O_5), and the Cl/P_2O_5 ratios become larger and variable. It is concluded that, in the andesite and the other rocks, chlorine is present in minerals other than apatite. Even in the two basalts of lowest phosphorus content, Cl/P_2O_5 is about 0.015, or somewhat greater than in the other four basalts. The impression is gained that the amount of apatite is not great enough to hold all the chlorine, and some of it is present in other forms.

The central Texas granites (Table 11) described by Goldich (1941) are fairly uniform in chlorine, except for the Petrick granite. The reason for the higher chlorine content in the Petrick is not evident from the chemical analysis.

TABLE 12.—CHLORINE IN COMPONENT MINERALS OF SOME IGNEOUS ROCKS
In per cent

Rock	Mineral	Percentage of rock	Cl in mineral	Absolute percentage Cl contributed to rock	Cl in rock by direct analysis
Quartz monzonite, Rockville, Minn.	Microcline	35	0.014	0.005	0.032
	Plagioclase (An 28)	35	0.028	0.010	
	Biotite*	7	0.18	0.013	
	Quartz		0.002	<0.001	
				0.029 total	
Hornblende granite, Stearns Co., Minn.	Hornblende (+biotite)	7	0.18	0.013	0.020
	Feldspar (mostly K)	~80	0.014	0.011	
	Quartz	~10	(0.00)	0.00	
				0.024 total	
Granite, Devils Slide, N.H.	Feldspar (chiefly perthite)	70	N.d.†	—	0.017
	Quartz	25	N.d.	—	
	Amphibole (sodic)	5	0.28	0.014	
Gabbro, Duluth, Minn., Fairmount Park, Kingston Creek at R.R.	Nonmagnetic heavy fraction‡	~15	0.025	0.004	0.012
	Plagioclase	~80	0.006	0.005	
				0.009 total	
Gabbro, Duluth, Minn.; along boulevard, above cement mill	Nonmagnetic heavy fraction§	~20	0.068	~0.015	0.022
Augite-syenite, Cook Co., Minn.	Nonmagnetic heavy fraction¶	~20	0.085	~0.02	0.041

* Includes 1% hornblende, traces of zircon and apatite.

† Not determined.

‡ Frequency percentage: pyroxene, 77; magnetite-ilmenite, 18; olivine, 3; plagioclase, 2; apatite, <1. Microscopical analysis by S. S. Goldich.

§ Frequency percentage: pyroxene, 11; amphibole, 41; epidote, 20; magnetite-ilmenite, 17; apatite, 3; plagioclase, 7; biotite, 1. Microscopical analysis by S. S. Goldich.

¶ Frequency percentage: pyroxene, 31; amphibole, 14; epidote, 17; magnetite-ilmenite, 21; apatite, 9; aggregates (leucoxene, epidote, etc.), 8. Microscopical analysis by S. S. Goldich.

MANNER OF OCCURRENCE OF CHLORINE IN IGNEOUS ROCKS

Chlorine can occur in rocks in a number of ways:

(1) In minerals in which it is an essential or predominant constituent, as in sodalite or scapolite. Quantitatively of little importance

(2) Isomorphously replacing OH^- and $O^=$, as in micas and apatite

(3) In aqueous inclusions, as in feldspar and quartz

(4) In "solid" solution, as in glasses

Despite the large difference in the ionic radii (Goldschmidt values) of OH^- (1.33A) and Cl^- (1.81A), chlorine appears to substitute for hydroxyl (or oxygen) in such common rock minerals as biotite and amphibole, and to some extent in apatite. Determinations of chlorine in biotite are scarce, but the data available indicate that as much as 0.2 per cent is not uncommon. Biotite from a Minnesota quartz monzonite (Table 12) contains 0.18 per cent Cl, which is very similar to the percentage (0.20) found in the biotite of the Butte granite (quartz monzonite) as reported by Weed

(1899, p. 743). The latter biotite also contains 0.76 per cent F. Amphiboles[3] may contain even more chlorine than biotites. Amphibole from a New Hampshire granite (Table 12) contains 0.29 per cent Cl (and 0.73 per cent F), and amphibole from rapakivi granite of the Salmi area, Finland, contains 0.51 per cent Cl (and 1.06 per cent F) as reported by Sahama (1947, p. 159). Apatite from low-silica rocks may contain considerable chlorine (Kind, 1939, p. 75). However, even in such rocks, it may be doubted whether apatite can account for the major part of the chlorine, except when phosphorus is abnormally high, as in the west Texas basalts.

In ultramafic rocks chlorine probably occurs to a large extent in the serpentinization products, substituting for OH^-. In two samples of gabbro (Duluth) and one of augite syenite, the chlorine content of the bromoform-heavy fraction (from which magnetite had been largely removed) was higher than that of the rock as a whole (Table 12). One gabbro yielded a heavy fraction consisting of ~40 per cent amphibole, 20 per cent epidote and 10 per cent pyroxene, with a chlorine content of ~0.07 per cent, compared to 0.022 per cent in the rock. The heavy fraction of the other gabbro, consisting largely of pyroxene, contained 0.025 per cent Cl, compared to 0.012 per cent in the rock. From the somewhat roughly determined percentages of the ferromagnesian minerals in these samples it appears that they do not account for all the chlorine, and that the remainder is to be found largely in the feldspars. The gabbro containing amphibole and epidote is richer in chlorine than the one in which pyroxene is the predominant ferromagnesian mineral.

By no means can all the chlorine in a granitic rock always be accounted for by hydroxyl-containing minerals such as biotite. This is demonstrated by the quartz monzonite in Table 12, which contains about half its total chlorine in the feldspar. The form in which chlorine is present in the feldspar has not been established, but other evidence points to its presence in fluid inclusions. Faber (1941) made a special study of the composition of the liquid inclusions in igneous rocks (almost all granites) and pegmatites. He found chlorine to be an important constituent of the solutions in the cavities. The percentage of chlorine (referred to the rock) in the liquid inclusions of granites, as obtained by leaching the finely ground rock with water, ranged from 0.005 to 0.031 in 13 specimens, with an average of 0.015. Pegmatitic feldspar from two localities contained 0.008 and 0.006 per cent water-soluble Cl, and the associated quartz, 0.012 and 0.015 per cent water-soluble Cl. The total chlorine content of the samples is not given. The weight ratio Cl/H_2O in the pegmatites is about 0.07, and in the granites 0.07 for six Precambrian samples and 0.04+ for younger granites.

The writers made a few determinations of soluble chlorine in 100-mesh rock or mineral powders by washing with water or sodium nitrate solution at room temperature and titrating chloride potentiometrically in the filtrate. The following percentages were obtained (figures in parentheses denote total chlorine percentage): granite, Westerly, R. I., 0.002 (0.009); granite, Aitkin Co., Minn., 0.000 (0.018); granodiorite, St. Cloud, Minn., 0.002 (0.023); gabbro, Duluth, Minn., 0.004 (0.022); diabase, Grand Marais, Minn., 0.002 (0.011); diabase, Fairfax Co., Va., 0.006 (0.021); anorthosite, Minn., 0.000 (0.005); serpentine, San Francisco, Calif., 0.004 (0.041); slate, Cook Co., Minn., 0.000 (0.023); mica schist, Minn., 0.001 (0.008); perthite, Graniteville, Mo., 0.007 (0.058); calcite, Cook Co., Minn., 0.010 (0.020). Because the powders were not finely ground, the values for soluble chlorine are minimum values.

Significant amounts of water-soluble chlorine are present in the gabbro and the diabases, and possibly in two of the granitic rocks. Doubtless most of the chlorine in the calcite is present in liquid inclusions; much of the chlorine in the perthite may have the same form.

The chlorine content of various minerals is listed in Table 13. Small amounts of chlorine are of universal occurrence; larger amounts are, in general, limited to minerals containing OH, but the presence of hydroxyl in the lattice

[3] A remarkably high chlorine content (7.24%) is shown by the amphibole *dashkesanite* found in a skarn in Transcaucasia (Krutov, 1936). The associated igneous rocks contain up to 0.13% Cl.

is not a guarantee of the presence of much chlorine.

Escape of Chlorine from the Exterior Shell of the Earth

In an attempt to estimate the maximum average chlorine content of magmas near the earth's surface, we may imagine the chlorine in the sea to be added to a thin shell of rocks whose present chlorine content is likely to be similar to that of accessible igneous rocks, *i.e.* ~0.02 per cent. Arbitrarily take this shell to be the "ten-mile crust", equivalent to 6 km. of granitic rock and 10 km. of basalt over the whole of the earth's surface. The mass of this shell per square cm. of the earth's surface is roughly 5×10^6 grams. Adding the oceanic chlorine,[4] 5.4×10^3 g. Cl per sq. cm., to the assumed percentage now present in the rocks, we obtain

$$\frac{5.4 \times 10^3 \times 10^2}{5 \times 10^6} + 0.02 = 0.13 \text{ per cent Cl.}$$

On this basis the maximum percentage of chlorine which could have escaped from the assumed shell of origin is

$$\frac{0.13 - 0.02}{0.13} \times 100 = 85.$$

Doubtless this value is much too large. There is no reason for believing that all the oceanic chloride has had such a shallow origin.[5]

The chlorine content of glasses is usually high, and available data give an average of *ca.* 0.08 per cent. If this value is taken to represent the chlorine in the magma, and the chlorine in the

[4] Chlorine in sedimentary rocks (dispersed, in solution, and in salt beds) is assumed to be small compared to oceanic chlorine.

[5] In this argument it does not seem necessary to consider the possibility of oceanic chlorine originating independently of magma at an early period of earth history. Brown (in Kuiper, 1949, p. 265) has shown from a comparison of cosmic and terrestrial abundances of the rare gases that the process of earth formation was probably such that any substance existing at the time primarily in the gaseous state was not retained in appreciable amount. A similar view is expressed by Kuiper (1949, p. 311). This view does not, however, seem to preclude the retention of chlorine released in the crystallization of relatively cool crustal material in late pregeologic time.

Table 13.—Chlorine in Miscellaneous Minerals

In per cent

Mineral	Cl
Perthite (pegmatitic), Graniteville, Mo.	0.058
K-feldspar, from St. Cloud, Minn., red granite	0.014
K-feldspar (locality unknown)	0.013
K-feldspar (pegmatitic), Warman, Minn.	0.018
K-feldspar (pegmatitic), Kanabec Co., Minn.	0.005
Amazonite, Mont Anpanobe, Madagascar	0.012
Amazonite, Kragerö, Norway	0.008
Amazonite, Amelia Court House, Va.	0.003
Bytownite, Crystal Bay, Minn.	0.005
Enstatite, N. C.	0.011
Hypersthene, Nain, Labrador	0.022
Diopside, Hull, Quebec	0.034
Diallage, Sonoma Co., Calif.	0.009
Anthophyllite, Salem district, Madras Province, India	0.05
Tremolite, from marble of Grenville Series, Fowler, N. Y.	0.02
Tremolite, same locality as preceding	0.03
Amphibole (edenite), Salem district, Madras Province, India*	0.025
Amphibole (pargasite), Salem district, Madras Province, India†	0.05
Amphibole (femaghastingsite) Lake Bonaparte Quadrangle, N. Y.‡	0.57
Beryl, Graniteville, Mo.	0.033
Topaz, Va.	0.010
Allanite, Morton, Minn.	0.29
Clinozoisite (from anorthite gneiss), Madras Province, India	0.07
Prehnite, Cook Co., Minn.	0.048
Staurolite, Morrison Co., Minn.	0.053
Tourmaline (black), Black Hills, S. D.	0.006
Biotite, St. Lawrence Co., N. Y.	0.11
Phlogopite	0.06
Lepidolite, Black Hills, S. D.	0.031
Stilpnomelane, Baern, Bohemia	0.043
Calcite, Two Harbors, Minn.	0.02
Fluorite, England	0.017
Quartz (with liquid inclusions)	0.003
Magnetite-ilmenite (black sand, from decomposition of mafic rocks), Cook Co., Minn.	0.014

* From anorthite-hornblende gneiss.

† Chrome amphibole from chromite amphibolite.

‡ From hornblende-quartz syenite gneiss.

crystallized rock amounts to ~0.02 per cent, a chlorine loss of 75 per cent is indicated. However, there may have been an enrichment of chlorine in the magma forming the glass, so that this figure is far from certain. Nevertheless, the generally high chlorine content of glasses is significant. Glasses containing 0.2–0.4 per cent chlorine are obviously derived from magmas much richer in chlorine than the average.

ALLOTMENT OF CHLORINE TO THE GRANITIC AND BASALTIC SHELL: If the chlorine in the sea is assumed to have come from the ~60 km. exterior shell of granitic material (6 km., reduced to total earth area) and basalt (54 km.), we have the relation:

$$\text{Mass of shell} \times \{\text{Cl}(\%)_{\text{shell, original}} - \text{Cl}(\%)_{\text{shell, present}}\} = 100 \times \text{mass of Cl in ocean.}$$

On the basis of masses (grams) per square cm. of the earth's surface:

$$\text{Cl}(\%)_{\text{shell, original}} - \text{Cl}(\%)_{\text{shell, present}} = \frac{100 \times 5.4 \times 10^{3}}{1.8 \times 10^{7}} = 0.03.$$

We thus have the value of the difference between two unknown quantities. To find either, assumptions must be made, and at best the conclusions are speculative. However, certain limits can be assigned which are of interest.

The lower limit of $\text{Cl}(\%)_{\text{shell, original}}$ is obtained by putting the present average Cl content of the shell equal to the chlorine content of accessible rocks (~0.02 per cent). This gives

$$\text{Cl}_{\text{shell, original}} > 0.03 + 0.02 > 0.05 \text{ per cent.}$$

It is reasonably certain that the present average Cl content of the shell is greater than ~0.02 per cent, because magma would be expected to lose chlorine in its ascent, and rocks crystallized near the surface have, on the whole, a greater opportunity to lose chlorine. It seems natural to consider $\text{Cl}_{\text{shell, original}}$ as the chlorine content of the primitive (predominantly basaltic) magma, which then would be >0.05 per cent (average).

The upper limit of the original chlorine in the shell (and therefore in the primitive magmas cannot be stated with much certainty. Reason) have already been given for the belief that the magmas from which the accessible rocks were formed did not contain more than 0.13 per cent Cl. Provisionally this value will be taken as the maximum of $\text{Cl}_{\text{shell, original}}$. The deep portion of the shell might have had a higher content than this, and the average would then be greater. However, the value of 0.13 per cent for the upper portion of the shell is very likely too great because the unlikely assumption has been made that all the oceanic chloride has originated in this shallow layer.

If $\text{Cl}_{\text{shell, original}}$ is accepted as <0.13 per cent:

$$\text{Cl}_{\text{shell, present}} < (0.13 - 0.03) < 0.1 \text{ per cent.}$$

The present average chlorine content of the shell being considered is a matter of importance, but unfortunately it seems to be impossible (at least on the basis of the present considerations) to specify it more closely than as lying in the range 0.02 to 0.1 per cent. The lower limit seems unlikely if the major part of the chlorine liberated has actually originated in the 50+ km. basaltic shell. A present average content of 0.02 per cent would imply the loss of some 60 per cent of the original chlorine, 0.03 per cent having gone to the ocean. This loss is so great that, if it has taken place more or less uniformly during geologic time, Archean igneous rocks should show higher chlorine contents on the average than young rocks. Available data are inadequate for a conclusion, but there is no indication of such an age effect. Independence of the chlorine content of a rock with time can be explained by the accumulation of the greater part of oceanic chlorine in pre-geologic time, by origin at greater depths than here assumed, or by a larger supply in the shell in question. An upper present limit of ~0.1 per cent chlorine in the exterior shell (~60 km.) would mean that approximately one-fourth (*i.e.*, 0.03/0.13) of the original supply has escaped up to the present time.

The assumption that much of the oceanic chlorine was exhaled in pre-geologic time seems to run counter to the present rate of escape of chlorine from the crust. The annual production is unknown, but the impression obtained is that it is large, perhaps great enough to permit the accumulation of oceanic chloride in the span of 3×10^{9} years comprising geologic time.

Even if the present rate were known it might not represent the average rate during this long period. Nevertheless, it is of some interest to calculate the contribution of some volcanic, fumarolic, and hot-spring areas, for which data

TABLE 14.—ANNUAL CHLORINE YIELD OF SOME MAGMATIC SOURCES
In grams

	Cl produced per year	Percentage of total yearly requirement
Kilauea	3×10^{10}	0.3
Valley of Ten Thousand Smokes (1919)	1.3×10^{12}	13
Norris Geyser Basin, Yellowstone Park	1×10^{9}	0.01
Steamboat Springs, Nevada	3.5×10^{9}	0.035
Arima, Hyogo Prefecture, Japan*	1.5×10^{9}	0.015

* The water flow of this spring is conservatively estimated at 100 tons per day; Cl = 40 g. per liter.

are available. The yearly production of Cl required to yield the chloride of the ocean in 3×10^9 years is $2.8 \times 10^{22}/3 \times 10^9$ g. or approximately 1×10^{13} g. In Table 14 the percentage of the yearly requirement furnished by certain areas is given. Most of the chlorine liberated is probably juvenile. When it is considered that the number of active volcanoes alone has been reported as 300 to 400, the world-wide production of chlorine at the present time must be a considerable fraction of 10^{13} g. per year.

These considerations do not lead to any certain conclusions, but suggest a deep source for some of the chlorine which has escaped from the crust. It has recently been shown (Kulp, 1951, p. 327), on the basis of certain assumptions, that 14 per cent of A^{40} produced from K^{40} during geologic time ($\sim 3 \times 10^9$ years) in the silicate portion of the earth has found its way into the atmosphere. If this value is approximately correct, some of the argon must have originated below the 60 km. exterior shell. By assuming an escape factor for water similar to that for argon, Kulp demonstrates that the ocean could have been derived from emanations from the earth's interior during geologic time. It would then be probable that chlorine in the ocean had a similar genesis, since the escaping tendencies of chlorine and water are similar.

TABLE 15.—ESCAPE RATIOS FOR CERTAIN ELEMENTS

Element	Ratio Amount escaped (g./cm.²)/ Amount in exterior shell (g./cm.²)
Cl	$5.4 \times 10^{9}/2.7 \times 10^{9} = 2.0$
H(as H_2O)	$3.1 \times 10^{4}/2.7 \times 10^{4} = 1.2$
B	$25/50 = 0.5$
C	$2 \times 10^{9}/5.4 \times 10^{9} = 0.4$*
N	$8 \times 10^{9}/8 \times 10^{9} = 1$
S	$5 \times 10^{9}/1 \times 10^{4} = 0.05$
Ne	$1.3 \times 10^{-3}/1.3 \times 10^{-3} = 10$
A	$12.9/0.7 = 16$
Cu	$8/1.2 \times 10^{3} < 0.01$†
Zn	$18/2 \times 10^{3} < 0.01$†

* Probably too low; the average value 0.10 per cent reported for CO_2 (0.03 per cent C) in igneous rocks is likely to be too high because of the presence of carbonate introduced by weathering.

† If a correction is applied for metal introduced into sedimentary rocks by weathering of igneous rocks, an escape ratio of ~ 0.002 is obtained.

Rubey (1951) has considered in detail the geologic history of sea water with special reference to the source of volatile constituents such as water and carbon dioxide. He concludes that these are most likely derived from plutonic gases released in the final crystallization of magmas. Hot springs are considered to be the principal channels of escape of the volatiles. To explain the continuous supply of volatiles, production of new magmas by local selective fusion of subcrustal rocks to a depth of several hundred kilometers is postulated. These magmas (granitic and hydrous) rise, slowly crystallize near the surface, and release their volatiles.

There would be no difficulty in accounting for a steady supply of liberated chlorine through the span of geologic time on the basis of a hypothesis of this type. If the depth of the shell that can be drawn upon is taken as 700 km. and its average chlorine content as little as 0.02 per cent, oceanic chlorine would represent less than 10 per cent of the original supply.

ESCAPING TENDENCIES OF CHLORINE AND OTHER ELEMENTS COMPARED: A measure of the tendency of elements to escape from the

earth's crust is taken to be given by the ratio of mass of element escaped (now in atmosphere, hydrosphere and sediments) to mass of element remaining in granitic and basaltic shell. The arbitrary shell chosen as a basis for comparison is again the 60 km. exterior shell. The content of the elements in the basaltic portion of this shell is of predominant importance and is assumed to be given by the content of accessible basaltic or basic rocks or, lacking such data, simply by igneous rocks. Such an assumption may be far from the truth. Although the absolute values of the ratios can thus have little significance, the comparative values should be a rough measure of the atmophilic and hydrophilic character of elements under the conditions existing in a relatively shallow earth layer. In Table 15 the ratios are given for chlorine (basalt = 0.015 per cent) and most of the other atmophilic and hydrophilic elements for which data are available. For comparison with these, copper and zinc have been included. For most of these elements it is not necessary to apply a correction for the amount of the element liberated by the weathering of igneous rocks.

These elements (excluding copper and zinc as not being markedly hydrophilic and still less atmophilic) fall into three groups on the basis of their escape ratios:

Ne, A	10–15	
Cl, H, B, C, N	~0.5–2	
S		~ 0.05

This grouping is in general accord with expectations (with the possible exception of N whose ratio seems low): Ne and A are gases and form no compounds; Cl, H, B, and C can be chemically bound in minerals, but they also form compounds of pronounced volatility and water solubility under the existing conditions; the chalcophilic character of S exerts a restraining effect on its atmophilic and hydrophilic tendencies.

The escaping tendency of chlorine (as well as of the other elements) must be closely related to that of water, which usually is present in much larger amounts in magmas. The two would be expected to be comparable, but *a priori* it is difficult to say which should be greater. The ratio Cl/H_2O+ increases in the order basalt → common granites → high silica granites, so that in this differentiation series relatively more water than chlorine has escaped. However, the two may have escaped in relatively more comparable amounts in crystallization of basalt.

The water content of basaltic magma has been estimated as 4 to 6 per cent (Phemister, 1934, p. 19–36; Gilluly, 1937, p. 440–41; Daly, 1944, p. 1379; *cf.* Ingerson, 1950, p. 810). Taking 5 per cent as a fair average for water, and, using the limits >0.05 and <0.13 per cent for average chlorine in the assumed primitive, predominantly basaltic magma, we obtain the limits >0.01 and <0.03 for Cl/H_2O. This result is similar to the ratio in the sea (0.019), and it appears that chlorine and water have been exhaled in proportionately like amounts. The gases from the Halemaumau, Hawaii, lava lake are reported to have an average weight ratio Cl/H_2O = 0.012. In the waters of some hot springs Cl/H_2O may be much higher than in the ocean. It is not unexpected to find the ratio lower in most hot-spring water than in the ocean because of dilution with ground water.

A comparison of the concentration of chlorine in the exterior shell of the earth (as represented by crustal rocks and the hydrosphere) with that in the deep silicate interior (presumably as represented by silicate meteorites) would be of interest, even though conjectural. However, since the chlorine content of meteorites is not well known at present, it seems best to postpone such comparison.

General Geochemistry of Chlorine

Chlorine is an element of cosmopolitan character in its distribution in the geospheres. Present data, not so certain as desirable, indicate its presence in average amounts of some hundredths of a per cent in both iron and silicate meteorites, or in roughly the same concentration as in igneous rocks (0.02 per cent). Chlorine may therefore be called an element of both siderophilic and lithophilic character. Moreover, the high solubility of most of its simple compounds in water and the volatility of some (particularly hydrogen chloride) place it among the hydrophilic and

atmophilic elements and explain its concentration in the sea.

It is perhaps not surprising to find chlorine in the metal phase of meteorites and in terrestrial iron (as in the Ovifak iron of Greenland). As a strongly electro-negative element it would be expected to combine with iron and be associated with it, just as do phosphorus and sulfur, the two elements[6] preceding it horizontally in the periodic table (Table 16).

TABLE 16.—GENERAL GEOCHEMICAL PROPERTIES OF ELEMENTS 15–18 AND 9

Atomic No.	15 P	16 S	17 Cl	18 A	9 F
Siderophilic..	+	+	+	−	−
Lithophilic...	+	(+)	+	−	+
Hydrophilic..	(+)	+	+	(+)	(+)
Atmophilic...	−	(+)	(+)	+	(+)

+ Pronounced.
(+) Subordinate or weak.
− Negligible.

Most elements, particularly the metals, are lithophilic to the extent that they are oxyphilic. Chlorine, however, has practically no oxyphilic properties. In a silicate melt it exists as a chloride or chloride ion. When crystallization occurs it is compelled to form a chloride (either a simple chloride or a compound containing chloride as an essential constituent) or to substitute for a negative ion of comparable size ($O^=$, OH^-). The manner of occurrence of chlorine in stony meteorites is obscure. Chlorapatite is said to be present in some meteorites, but apparently phosphorus is usually present as merrillite, a sodium calcium phosphate. Moreover, the Cl/P ratio in some meteorites is such that apatite could not account for all the chlorine. The possibility does not seem to be excluded that chlorine is present as a more or less disseminated chloride, such as ferrous chloride. Merrill speaks of the discoloration of stony meteorites caused by the oxidation of lawrencite.

In the upper levels of the earth, under more hydrous conditions, replacement of OH^- by Cl^- becomes important. Because of the difference in size between OH^- and Cl^-, there is a certain reluctance for the substitution to take place, and, because of its hydrophilic character, chlorine becomes associated with water. At and near the earth's surface the atmophilic properties of the element come into play. Much of the chlorine escapes as hydrogen chloride formed by hydrolysis at high temperatures:

$$Cl^- + H_2O \rightleftarrows HCl + OH^-$$

The geochemical behavior of chlorine differs in a number of respects from that of fluorine. The siderophilic tendency of fluorine appears to be very weak; the fluorine analog of lawrencite has not been observed in meteorites. The antipathy of fluorine toward the iron phase is understandable from its special affinity for silicon, in which respect it differs markedly from chlorine. Available data indicate that in silicate meteorites fluorine occurs to a much smaller extent than does chlorine. The Noddacks (1934, p. 178) have reported 0.004% F in stony meteorites, which is about one-tenth that of chlorine. In the outer shell of the earth, chlorine and fluorine are on a more even footing, and fluorine probably exceeds chlorine in silicate rocks (*cf.* Barth, 1947, and Wasserstein, 1947). As already noted, fluorine shows a greater tendency than does chlorine to remain in rocks formed by crystallization of the magma. Such analyses as are available show less fluorine than chlorine in volcanic gases; chlorides are more abundant than fluorides or fluosilicates in volcanic sublimates and waters of magmatic origin. The greater ability of fluorine to replace $O^=$ and OH^- in minerals and the insolubility of calcium fluoride explain this behavior. The great store of chlorine in the ocean is further evidence of the greater escaping tendency of chlorine. Even if allowance is made for the fluorine in phosphatic sediments it seems certain that more chlorine than fluorine has been exhaled. The abundance of chlorine may exceed that of fluorine in the earth as a whole, because

[6] Nonmetals such as phosphorus, sulfur, and chlorine should perhaps be classified as pseudo-siderophilic elements in the sense that they do not form a solid solution in the iron (as do truly siderophilic metals) but, in the form of their compounds with iron, produce a separate phase. In the case of sulfur the formation of the separate phase (troilite) is of importance as leading to the segregation of chalcophilic elements therein, and the distinction has always been made.

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of the increase in the ratio Cl/F in the deeper layers.

The cosmic abundances of fluorine and chlorine, based on planetary nebulae data, appear to be similar. Brown (1949, p. 628) accepts the values 90 and 170 as the atomic abundances of fluorine and chlorine referred to silicon as 10000.

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SKETCH MAP SHOWING OUTER WISCONSIN DRIFT BORDER (IOWAN-TAZEWELL COMPLEX?) AND SUGGESTED INNER DRIFT BORDER (CARY?) BETWEEN CENTRAL NEW YORK AND THE MASSACHUSETTS COAST

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PROBABLE WISCONSIN SUBSTAGES AND LATE-WISCONSIN EVENTS IN NORTHEASTERN UNITED STATES AND SOUTHEASTERN CANADA

BY RICHARD FOSTER FLINT

ABSTRACT

A variety of evidence suggests that the borders of the Cary and Mankato drift sheets cross eastern New York and New England, that there were two and perhaps three late Wisconsin marine invasions of the St. Lawrence lowland, and that there exist logical eastern correlatives of the low-water phases in the glacial upper Great Lakes, in the form of till-covered marine deposits, the abandoned, drift-filled Niagara gorge, and the North Bay–Ottawa River lake-outlet channel.

The suggested events from the Cary maximum to the present are assembled in a correlation chart for comparison and criticism. Although the scheme proposed is suggestive only, it is consistent with ecology implied by fossils, with the dates and amplitudes of known crustal movements, and with the radium- and radiocarbon dates of the dated materials collected from the region.

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Introduction

The Glacial Map of North America (Flint and others, 1945) shows the Wisconsin drift in eastern New York, New England, the Maritime Provinces, and Quebec as "Wisconsin, undifferentiated." The implied lack of knowledge reflects the great relief, steep slopes, resistant, quartz-rich bedrocks, and thin glacial drift that characterize much of this region, furnishing conditions unfavorable for unhindered flow of glacier ice and for the creation and preservation of stratigraphic sections including more than one drift sheet. The absence of end moraines in this region is almost proverbial; it contrasts strongly with the intricate pattern of moraines in the Great Lakes region and reflects differences in the response of the ice sheet, at least in detail, to the fluctuations of climate during the Wisconsin age. Indeed, some of the fragmentary features earlier accepted as end moraines were later held by Goldthwait (1938) not to be end moraines at all. A large proportion of the drift is stratified and is concentrated in the valleys, regardless of valley orientation.

Despite the marked differences between the Atlantic Slope and the Middle West, an array of detailed observations of glacial stratigraphy and morphology within the former region has accumulated. The discussion that follows is an attempt to arrange these observations into a rational sequence, and to infer from them the positions of the drift borders of the Wisconsin substages and the lacustrine and marine features related to them. The correlations suggested are no more than a framework to focus attention on the problem, stimulate further study, and provide a concrete target for attack. Only with the setting up of such targets from time to time can we progress toward the truth.

Specifically, it will be proposed:

(1) that the borders of the Cary and Mankato drift sheets cross eastern New York and New England;
(2) that north of the periphery of the Wisconsin drift there is evidence of two major readvances of the ice-sheet margin (Cary and Mankato) and, in addition, lesser readvances postdating the Mankato maximum.
(3) that along the New England coast there are two sets of glacial-marine deposits, respectively of late Tazewell and late Cary age;
(4) that there were at least two and probably three marine invasions of the St. Lawrence lowland, the three corresponding respectively to Two Creeks, late Mankato, and perhaps Thermal Maximum;
(5) that there exist logical eastern correlatives of the low-water phases of the western Great Lakes in the form of till-covered marine deposits, the abandoned, drift-filled Niagara gorge, and the North Bay-Ottawa River channel; and
(6) that the scheme proposed is consistent with ecology implied by fossils, with the dates and amplitudes of crustal uplifts, and with the radium- and radiocarbon dates of the few dated materials collected from the region.

Much detailed field study will have to be accomplished before the validity of the correlation proposed can be established. However, the scheme explains features hitherto ignored or considered unrelated. At least it will focus attention on late-Wisconsin features in northeastern North America and stimulate further inquiry.

For convenience, the older Wisconsin features are discussed by sectors of the glaciated region; the younger features are discussed in stratigraphic and chronologic sequence. No attempt is made to refer to the entire literature on this problem. Only critical references are cited, most of them pertaining to relatively recent investigations.

Acknowledgment

I wish to express my thanks to E. S. Deevey, Jr., C. D. Holmes, Paul MacClintock, and G. M. Stanley for thoughtful reading and discussion of the manuscript.

Long Island-Connecticut Sector

Outer Belt

In the Connecticut-Long Island sector an outer belt of territory some 50 miles wide is bounded on the south by two closely related end moraines. The moraines constitute the twin peninsulas at the eastern end of Long Island and continue eastward into Rhode Island. In western Long Island the outer, Ronkonkoma moraine is overlapped by the inner, Harbo

Hill, moraine. Three glacial advances are recorded—a Ronkonkoma advance, a later, Harbor Hill, advance, and a still later readvance that in places overrode the Harbor Hill moraine. Apparently the time intervals between the advances were short (Fleming, 1935).

The outer moraines, combined into a single one, cross from Long Island to Staten Island, forming the Wisconsin drift border. The border, with or without end moraine, traverses New Jersey, Pennsylvania, and southwestern New York State.

At places on the mainland, in Connecticut but still within the 50-mile outer belt, ventifacts and evidence of frost disturbance of the drift occur (Denny, 1936). Such features are expectable results of the cold and windy conditions at the maximum of a major glaciation. Search has failed to reveal evidence of readvance during the deglaciation of the belt, except for a minor readvance at Montowese on the Quinnipiac River (Flint, 1934, p. 84–85).

Bodies of varved clay and silt, accumulated during the uncovering of this belt, occupy the lower Quinnipiac valley (Flint, 1934) and the vicinities of Middletown, Berlin, and New Britain (Flint, 1933), as well as Bloomfield (G. A. Bourbeau, unpublished). In places they are overlain by till.

Inner Belt

Along a line about 50 miles inside the Ronkonkoma moraine, there is evidence of a pronounced stand of a glacier margin, following a conspicuous readvance. In the Connecticut River drainage the readvance is marked by the occurrence of both till and ice-contact stratified drift overlying disturbed varved sediments throughout a belt more than 16 miles wide measured from north to south (Flint, 1933, amplified by later unpublished data). The stand following the readvance is marked by a bulky mass of stratified drift that formerly blocked the Connecticut River, but that is now trenched longitudinally by it. The mass is part kame terrace, part outwash valley train. It has an ice-contact proximal end in the vicinity of Rocky Hill and is deeply pitted with kettles as much as a mile in individual diameter, as far as Middletown, 8 miles downstream. Southeast of Middletown it continues as normal outwash, much eroded by the Connecticut River but still represented by terrace remnants. The headward part of this great mass has a maximum thickness of at least 300 feet. It extends westward across the Berlin basin and has its counterpart, only slightly less bulky, in the Quinnipiac Valley, reaching from Southington southward into Cheshire. Northward along both sides of the Connecticut Valley lowland, the sands and gravels continue as kame terraces, built during the thinning of the glacial lobe whose terminus has been described.

In the lower Quinnipiac valley both the altitudes and the internal character of erosion remnants of outwash sediments suggest that two valley trains are represented—an older, more massive, greatly dissected train and a younger, originally less massive, less dissected body.

The inference is offered that, following deglaciation a glacier reinvaded Connecticut, reaching as far south as Middletown and Cheshire, and that its meltwaters built a bulky body of sand and gravel upon and beyond its thin, wasting terminal zone.

Correlatives of these features have not yet been identified in the higher, more deeply dissected country of western Connecticut, but in the more open eastern part of the State several sections exposing a sandy till overlying a compact, weathered clay-rich till are reported (White, 1947). All lie north of the 50-mile outer belt.

A prolonged retreat followed the readvance to the vicinity of Middletown. This is established by the presence in the Connecticut Valley of lacustrine varved clay and silt, extending from the outwash head at Rocky Hill through northern Connecticut, Massachusetts, Vermont, and New Hampshire. These lake deposits in Connecticut and Massachusetts constitute a single body of sediment (Flint, 1933; Jahns and Willard, 1942); it is probable, although not established, that the deposits in Vermont and New Hampshire belong to the same lake. The outwash head acted as a high dam, and the water escaped at New Britain over a bedrock channel that detoured the dam. Thousands of varves have been counted in the sediments (Antevs, 1922); presumably therefore a period of thousands of years, following retreat from the outwash head, was marked by quiet sedi-

mentation in the enduring lake, with little if any evidence of further readvance.

Probable Correlations

In the Connecticut-Long Island sector the 50-mile outer belt with its pronounced end moraines, its local ventifacts and frost disturbance, and its weathered surface is believed to be the early Wisconsin drift, the Iowan-Tazewell complex. It is conceivable, even, that the three glacial advances recorded on Long Island represent, respectively, Farmdale (*see* Leonard, 1951, p. 330), Iowan, and Tazewell, for the Farmdale episode is recorded in both Illinois and Kansas, and the close relation of the moraines in time and in areal position favors such a correlation. This, however, is mere speculation; the early Wisconsin correlation of the complex as a whole is a probability.

The drift north of the 50-mile outer belt, with its evidence of readvance over an earlier drift, its sections showing two tills, and its bulky outwash, is believed to be the Cary substage, the first of the later Wisconsin drift sheets.

Massachusetts-Coastal New Hampshire Sector

Outer Belt

Like the Connecticut-Long Island sector, the sector in Massachusetts and coastal New Hampshire possesses two distinct belts of terrain. The outer belt includes Cape Cod and the islands and extends inland to the vicinity of Taunton. In the neighborhood of Boston its inner limit passes northeastward beneath the sea. It includes the continuations of the Ronkonkoma moraine (the end moraine of Martha's Vineyard and Nantucket islands) and the Harbor Hill moraine (the Buzzards Bay and Sandwich moraines on Cape Cod). Each represents a glacial advance, and a third advance is recorded by a smaller and later moraine (Mather, Goldthwait, and Thiesmeyer, 1940, p. 14). The three advances evident on Long Island are therefore evident here also. The multiple tills reported by Sayles and Knox (1943) appear to be referable to these advances. There is abundant evidence that stagnant ice related to the outer moraine persisted throughout the building of outwash from the inner moraines, and hence that the intervening intervals were short. Ventifacts, present in coastal Connecticut, occur here in far greater abundance, suggesting again the glacial-maximum conditions evident at the Iowan and Tazewell drift borders in Iowa and in South Dakota (Flint, 1950).

Drift belonging to this outer belt is described further in the next section, for it underlies younger drift in the inner belt.

Inner Belt

In the vicinity of Boston the stratigraphic section is more complex. As in Connecticut there is clear evidence of two glacial advances, but, thanks to the abundance of subsurface data from the metropolitan area, the details are far better known and have been assembled and interpreted in an excellent study by Judson (1949).

A compact, clay-rich till, locally built into drumlins, is overlain by a thick deposit of marine clay. The till and clay are weathered and in places deeply channeled. Overlying these weathered members are a loose, sandy younger till, gravelly and sandy outwash, in places a frost-disturbed zone, and, finally, marine silt and tidal-marsh peat. In many localities north and west of Boston and locally in the southeast part of Boston a loose sandy till overlies a compact clay-rich till. Judson believed these tills to be the same as the two tills described by him, and ascribed substage value to each, on the basis of the intervening record of weathering and channeling. He inferred a marine submergence during the deglaciation following the deposition of each of the two tills. The marine deposits in the Boston area lie mostly below sea level. The explanation of their absence in the Cape Cod region probably lies in northwestward upwarping during deglaciation; this would carry the marine deposits well below sea level in the vicinity of Cape Cod. Probably the brackish-water deposits inferred by Hyyppä (1939) from a study of diatoms in the region south of Boston belong to the higher substage. The brackish condition may be transitional from the marine area at and northeast of Boston, more deeply depressed by the ice sheet, to the area southwest

of Massachusetts, in which no evidence of glacial-marine conditions at and above today's sea level have been found.

The two bodies of marine sediment identified in the Boston area have long been recognized in northeastern Massachusetts and New Hampshire (*e.g.*, Clapp, 1907). The older antedates the last glaciation of the region; the younger postdates it.

Maine-New Brunswick Sector

In coastal Maine the Pleistocene section is closely similar to that in adjacent parts of New Hampshire and Massachusetts. It may be summarized thus from Perkins (Leavitt and Perkins, 1935, p. 189–191, 198–204):

4. Marine sediments with same fauna as (2).
3. Drift, including eskers. (Ice flowed south.)
2. Marine sediments with boreal invertebrate fauna, the sediments mashed and deformed as though by bergs.
1. Drift, including eskers. (Ice flowed southeast.)

The marine deposits have yielded, in addition to invertebrates, fishes, a sea bird (murre), two genera of seals, Atlantic walrus, and a whale.

There can be little doubt that the two drift sheets (*see also* Clapp, 1907) are the same as those recorded farther south, and that the later of the two represents the latest glaciation of coastal Maine, inasmuch as the marine deposits overlying it have not been overridden. That the older marine deposits, reaching higher altitudes, have been overridden partly explains the difficulty repeatedly experienced in the attempt to recognize strandlines related to them.

In New Brunswick a similar sequence is reported (Chalmers, 1896, p. 36–38). At Negrotown Point, St. John, the following section was exposed in 1894:

6. Marine clay and sand, forming distinct terraces [not later overridden]
5. Till
4. Stratified drift
3. Till with marine invertebrate shells
2. Laminated clay with marine invertebrate shells and erratic boulders
1. Till?

Although evidence of inter-till weathering is not reported, this section resembles the Maine sequence so closely that provisional correlation of the two sequences is justified.

The position of the submerged border of the Wisconsin drift in this sector is not known, although sea-floor contouring by Murray (1947) suggests that it extends at least as far southeast as Lat. 42°30′, Long. 68°00′. The conspicuous morainelike ridges near Lat. 43°00′, Long. 68°30′ (Chadwick, 1949) may possibly represent the outer limit of the later glaciation of the Boston area.

Correlation of Substages in Coastal New England

I suggest that throughout coastal New England the Wisconsin stratigraphy as now understood is consistent, and that it represents a lower element comprising various glacial advances separated by short intervals, and an upper element separated from the lower by weathering and dissection, each element being accompanied by a marine facies. I submit that the two elements are, respectively, the Farmdale-Iowan-Tazewell complex and the Cary substage, and that the Cary drift border passes east-northeasterly through Southington-Cheshire and Middletown in Connecticut, and disappears beneath the sea in the general vicinity of Boston, possibly being continued in submerged moraines on the floor of the Gulf of Maine. I submit further that the weathering and erosion of the earlier drift and marine deposits represent the Tazewell-Cary interval, also called the Brady[1] interval (*see* Leonard, 1951, p. 325) in the Middle West. Supporting evidence is as follows:

1. The outer complex is continuous with the outer Wisconsin drift from Long Island westward across the Appalachian region into the Central Lowland, where it is known to include the Tazewell substage.

2. Only one period of weathering (other than very slight leaching of calcareous loess) and dissection within the Wisconsin age is known: that between the Tazewell and Cary sub-ages. No exposures showing intra-Wisconsin weathering in New England, other than those mentioned in this paper, are known to me.

[1] In the reference cited the word form "Bradyan" is used. I have here written it in the standard form of general usage.

3. The suggested Cary drift border, 50 miles inside the older drift border, parallels the latter, parallels the trend of the marine isobases, and is normal to the striations.

4. The tentative age—18,000 years—obtained by Urry (1948) from the radium content of a clay-silt varve deposited near Hartford in the Connecticut valley lake during the glacial retreat following the readvance that culminated near Middletown is consistent with a Cary date as inferred from radiocarbon measurements. This varve, no. 3700 of Antevs (1922), is underlain by 700 measured varves and by at least 150 feet more of the same varved sediment. If we assume an average thickness of 1 inch per varve, the number of varves antedating varve 3700 is at least 2500. On this estimate the glacier terminus stood at the Rocky Hill outwash head, south of Hartford, at least 20,500 years ago, and at its extreme limit near Middletown, still earlier. The two early Cary samples from the Middle West that have been radiocarbon-dated (Libby, 1952, nos. 465, 508) have yielded ages of more than 15,000 years and more than 17,000 years, respectively.

5. The Cary drift border must lie southeast of Aroostook County, Maine. Pollen profiles from peat and limnic sediments in that vicinity exhibit a typical Alleröd- and younger sequence, the Alleröd being near the base of the section, which overlies the youngest till in the district (Deevey, 1951). As Alleröd is equivalent to the Two Creeks horizon (Cary-Mankato interval) (Flint and Deevey, 1951), the till beneath it can be no younger than Cary.

This evidence is far from establishing proof, but it ranges from permissive to strongly suggestive. The probability seems fair that the correlation proposed is valid.

Too little is known about the Wisconsin drift of Nova Scotia and Newfoundland to justify even such tentative inferences as the foregoing. However, the peat overlain by till reported by J. W. Dawson (1872, p. 178) on River Inhabitants near the southern coast of Cape Breton Island, if sampled and radiocarbon-dated, should indicate whether the latest glacial advance in that district is the Cary or the Mankato. The three successive upwarped marine water planes mapped in Newfoundland (Flint, 1940) probably are related to times of deglaciation, and when more information about the drift sheets has been accumulated, the warped marine horizons can be dated with respect to them.

Hudson Valley-Catskill Mountains Sector

Outer Features and Hudson Valley Lakes

The Long Island end-moraine belt continues across the Hudson River. North of and nearly contemporaneous with that belt is a body of varved silt and clay extending from the moraine northward up the river as far as Haverstraw, New York. The sedimentary body, deposited in glacial Lake Hackensack, rises northward, presumably in consequence of crustal uplift accompanying deglaciation.

The Lake Hackensack deposits reach an altitude of more than 100 feet (Woodworth, 1905, p. 100), show no evidence of overriding (Antevs, 1928, p. 112, 115), and are commonly covered with outwash sand and gravel (Peet, 1904, p. 419). The rising sea level of recent time has submerged the lower Hudson; hence the position and character of the lake outlet are unknown.

If correlation on a basis of varve measurements is accepted, Lake Hackensack was contemporaneous with the lake in the lower Quinnipiac Valley in Connecticut (Antevs, 1928, p. 114).

Farther north up the Hudson, another body of varved clay and silt extends from the vicinity of Glens Falls, New York, southward to the gap or gorge by which the Hudson traverses the Highlands between Newburgh and Peekskill. South of Newburgh this lacustrine body passes below sea level (the river surface). Northward it rises to an altitude of nearly 400 feet near Glens Falls. Woodworth (1905) ascribed the sediments to a glacial lake which he called Lake Albany. He believed the lake surface was controlled by a bedrock spillway south of Newburgh, now submerged beneath the tidal river. The dam and spillway may well have been similar to those which controlled the lake in the Connecticut Valley.

The lacustrine sediments exhibit no sign of glacial overriding between Newburgh and a point a few miles south of Glens Falls (although they do so farther north). They are not covered with outwash but are the latest glacial deposit in

the valley. Antevs (1928, p. 114) correlated Lake Albany varves near Newburgh with varves in the Connecticut Valley lake near Hartford, but was unable to find a connection between the Lake Albany varves and those of Lake Hackensack.

By vertical position and by varve correlation, the Lake Hackensack sediments and the Lake Albany sediments appear to be distinct, separate sedimentary bodies. Furthermore the fact that the Hackensack deposits are covered in places with outwash, whereas the Lake Albany deposits are not, suggests the presence of a glacier terminus south of Lake Albany but north of Lake Hackensack. The lack of evidence of overriding in both sedimentary bodies makes it improbable that the outer limit of a glacial advance could be located elsewhere than in the narrow gorge of the Highlands crossing—the area between the two lacustrine bodies—but unfortunately scour by the Hudson has removed whatever Pleistocene deposits may formerly have existed here above the surface of the present river.

The northern end of this critical zone is just 50 miles north of the outer moraine where that feature is cut by the Hudson River. This is the approximate width of the outer belt of drift in New England.

I suggest therefore that the Lake Hackensack deposits date from the retreat of the Tazewell glacier margin, and that the Lake Albany deposits date from the retreat of the Cary glacier margin, following a readvance that brought the Hudson lobe of the Cary ice sheet to a position between Newburgh and Peekskill. This correlation receives further support from relationships in the Catskill Mountains district, set forth in the following section.

A possible objection to the correlation of Lake Albany with the Connecticut Valley lake exists in a discrepancy between the estimated rates of northward rise, caused by crustal upwarping, of the two bodies of sediment. The rate of rise for the Connecticut Valley body, based mainly on deltas, is 4.2 feet per mile (Jahns and Willard, 1942, p. 274), whereas the rate for the Hudson body, based merely on the difference in altitude between the top of the sediments at Glens Falls and those at Newburgh, is 2.25 feet per mile. The two valleys are parallel and hence should give nearly equal figures. Field measurements in the Hudson Valley, however, have not been made, and until this has been done the reality of the discrepancy may be fairly questioned.

The continuity of the Lake Albany varved sediments precludes the existence of a Cary readvance at any latitude between that of Newburgh and that of Glens Falls. Although evidence of readvance near Glens Falls exists, that readvance must be post-Cary. If it were as early as Cary it would not fit the known sequence of glacial lakes and marine invasions shown on Plate 3.

The probability follows that Lake Albany must have been destroyed by draining in Cary time rather than in Mankato time, because the low-water phase in the Great Lakes and the correlative marine phase in the St. Lawrence lowland, which marked the Cary-Mankato interval, demand deglaciation of the Hudson-Champlain depression. Hence the eastward discharge down the Mohawk Valley, which built the massive Albany delta into Lake Albany, must have been the discharge through the Syracuse channels identified with the Whittlesey and Vanuxem (late Cary) lakes (Pl. 3) rather than one of the similar discharges of Mankato date.

Furthermore, the occurrence of a tundra-type reindeer in the Albany delta deposits (Fisher and Orstrom, 1952) implies the near proximity of the ice-sheet margin. This is more consistent with a late-Cary date than with a Mankato date for the delta, and is not consistent with the Lake Iroquois date assumed by Fairchild and cited by Fisher and Orstrom.

Catskill Mountains District

Rich (1935) separated the southern part of the Catskill Mountains district into two unlike areas:

> "one to the north and east, in which moraine loops are abundant, sharp and fresh; and the other to the south and west where few moraines are found, where smooth, thick drift is the prevailing form of glacial deposit, where evidences of the erosive action of the ice are few, and where the topography and the weathering of boulders suggest noticeably greater age of the drift." (p. 130).

He (p. 121) defined the rather indistinct boundary of the younger drift, which he termed "late Wisconsin," as running from Delhi

sinuously southeastward to Curry, and suggested that it might continue southeastward through Phillipsport. He suggested further (p. 123) that during the later of the two glaciations the higher parts of the southern Catskills were occupied by local glaciers.

The differences mentioned by Rich seem consistent with the differences between the outer and inner belts of drift in coastal New England. If they are consistent, the two drifts in the Catskill Mountains district can be interpreted as possessing substage value. Furthermore the outer limit of the younger drift as traced by Rich agrees satisfactorily with the inferred glacier terminus in the Hudson Valley at the Highlands crossing. In fact the agreement between the Hudson Valley and the Catskill features is good enough to justify the hypothesis that the younger drift in the Catskill Mountains district is the Cary drift, and that the subdued topography and the weathering of the older drift are a record of the Brady interval.

It is true that the outer belt of older drift in northern New Jersey, northeastern Pennsylvania, and southeastern New York is nearly twice as wide as the same belt in southern New England. I suggest that this difference is at least in part the result of the presence of the Catskill Mountains mass which, acting as a baffle, retarded the southwestward flow of the ice sheet in this sector.

Correlation of Substages in the New England, Hudson-Catskills, and Western New York Sectors

The outer belt of drift, extending from Cape Cod to the mouth of the Hudson, continues northwestward across New Jersey, Pennsylvania, and New York with a well-established outer limit, the position of which has long been known. This belt of drift appears to be a unit, belonging to a single substage, except for the outermost moraines on Long Island and the Cape Cod district, which may belong to a still earlier though closely related substage.

In the western New York sector this outer belt of drift was mapped by MacClintock and Apfel (1944) as the Olean drift and was believed by them (p. 1163) to be probably Tazewell. If the outer belt is continuous from western New York to New England, the evidence, in the latter sector, of weathering and dissection of that drift before the readvance responsible for the next younger drift strongly supports a Tazewell correlation.

The margin of the inner belt of drift extends from near Boston through central Connecticut, the Newburgh district, and the southwestern Catskills, ending westward at Delhi, New York, beyond which Rich did not trace it.

In western and central New York the border of a post-Olean drift was mapped by MacClintock and Apfel (1944) and was called by them the Binghamton drift. They noted (p. 1162) that whereas the Olean drift area is marked by numerous bulky fans on valley floors, testifying to considerable post-Olean erosion, the Binghamton drift lacks such features; hence the Olean-Binghamton interval was marked by substantial erosion. Although comparative lithology shows that the Binghamton drift was brought in from the north and northwest, and the Olean drift from the northeast (p. 1153-1154), the implied shift in the effective center of outflow of the glacier ice does not establish a considerable lapse of time, and hence does not enter into evidence. MacClintock and Apfel believed the Binghamton drift to be probably Cary.

The Binghamton drift border was traced by them eastward through Hornell, Elmira, Binghamton, and Oneonta. East of Oneonta they found that the lithologic distinction between the Olean drift and the Binghamton drift becomes uncertain (p. 1157), and they did not trace the border farther. Oneonta is just 15 miles northwest of Delhi, the northwesternmost point on the line separating a young little-eroded drift from an older, more eroded drift in the Catskill Mountains district. A continuation of this line as traced by Rich toward the northwest, would pass through or very close to Oneonta.

The extent of erosion separating the times of deposition of the two belts of drift, the near connection of the boundary traced west by Rich with that traced east by MacClintock and Apfel, the Cary date applied to the younger Catskill drift by analogy with the New England sequence, and the Cary date ascribed to the Binghamton drift by MacClintock and Apfel make a case for the following suggestion: The Cary drift includes the Binghamton drift, the

SKETCH MAP SHOWING SUGGESTED
CORRELATIONS IN EASTERN NE
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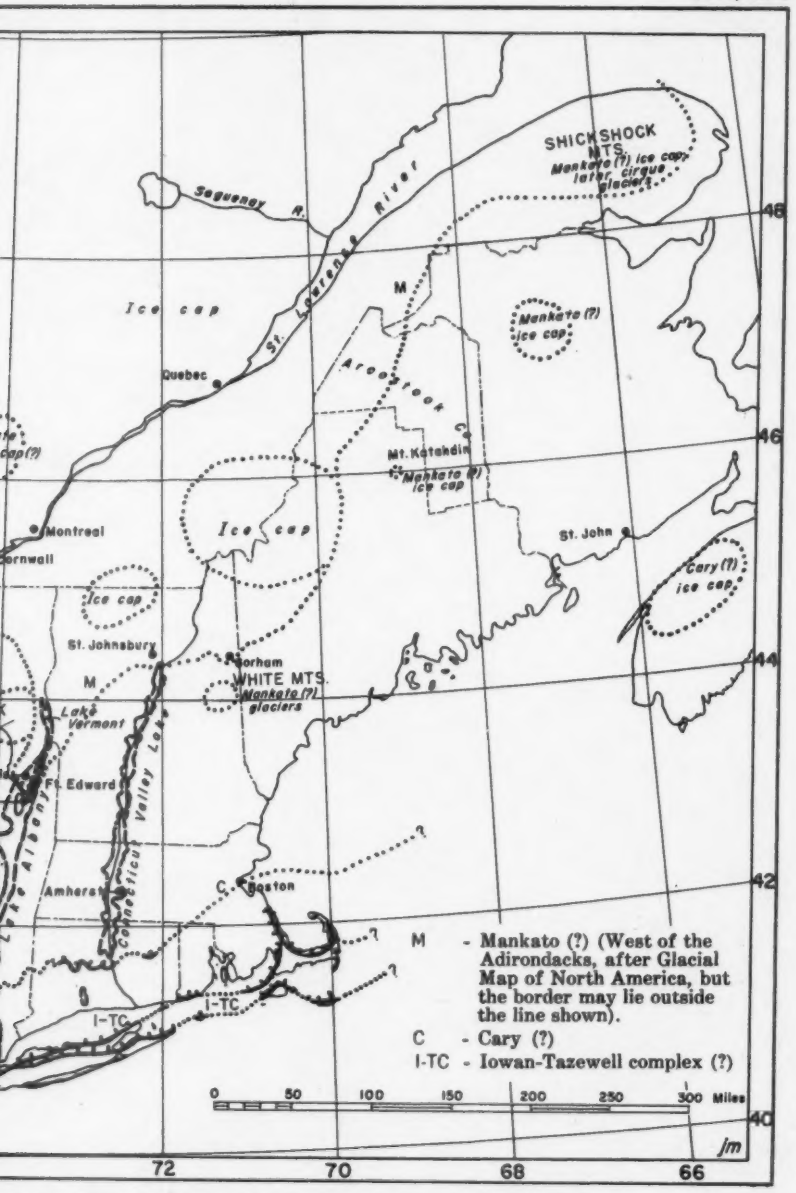

GESTED WISCONSIN DRIFT-BORDER
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younger drift in the Catskills, and the inner belt of drift in southern New England. The border of the single drift sheet, correlated on this basis, is reasonable topographically, departing from fair parallelism with the border of the older Wisconsin drift only in the Catskill sector.[2]

If this correlation should prove valid, the Valley Heads moraine, shown on the Glacial Map of North America as marking the Cary drift border, must postdate the Cary maximum. I know of no evidence that precludes such a reclassification of the Valley Heads moraine.

Two Creeks Interval

Position of Glacier Margin during Bowmanville Low-water Phase

Restudy of the relations of strandlines to end moraines in the Michigan and Huron basins (Bretz, 1951a; 1951b) has clarified considerably the events of the Two Creeks (Cary-Mankato) interval, already well known (*e.g.*, Flint and Deevey, 1951, p. 261–263) for its spruce bog, radiocarbon-dated at 11,400 ± 350 years. From various lines of evidence it appears that during that interval the margin of the Cary ice sheet retreated to a line at an unknown distance north of the Strait of Mackinac, lowering the water levels in the Michigan and Huron basins and permitting lake discharge toward the east. This was the Bowmanville low-water phase. In the Lake Michigan basin, the retreat amounted to at least 350 miles, measured from the Valparaiso moraine in northern Indiana, the Cary maximum, to the northern shore of the Strait of Mackinac. Following this retreat the glacier margin readvanced at least 275 miles to a position south of Milwaukee (Bretz, 1951b, p. 426). At the same times the ice margin in the New York State sector receded through an unknown distance north of the Syracuse region (as is known from the fact that the ice-dammed Lake Whittlesey ceased to exist, and drainage flowed east) and later readvanced at least as far south as Syracuse, closing the eastward drainage way and creating Lake Warren, which drained westward.

The extent of deglaciation in the New York sector is not known from direct evidence. However, from the near-parallelism of the Tazewell, Cary, and Mankato drift borders (including those suggested in the present argument) we can reason that except for differences imposed by relief features, the margin of the ice sheet during times of glacial minima should have been roughly parallel with the repeatedly similar margins at times of glacial maxima. It is, moreover, improbable that an event in the Lake Michigan region, clearly caused by important climatic fluctuation, was not strongly if not equally represented in the New York sector 400 miles to the east.

Let us assume a minimum Cary-Two Creeks retreat in the New York sector equal to the minimum retreat known to have occurred in the Lake Michigan basin, namely 350 miles. An equal retreat measured from the Binghamton (Cary?) drift border would give us a glacier margin, in Two Creeks time, standing in southwestern Quebec well north of the Ottawa River. Even if we assume a retreat of only half that amount, the glacier margin would have trended northeastward across southeastern Ontario, north of Kingston. In deducing these position we must remember that the 350-mile figure for the Lake Michigan region is a minimum distance, and that the retreat in that sector may have been considerably more.

In the first case the North Bay-Ottawa River outlet, used much later by the Nipissing Great Lakes, would have been uncovered and available as a drainage channel. In the second case the Trent Valley outlet, used later by Lake Algonquin, would have been uncovered. In either case, the St. Lawrence lowland probably would have been ice free and therefore open to marine invasion.

St. David Drift-filled Gorge

During the Bowmanville low-water phase (the lacustrine equivalent of the Two Creeks interval) the Michigan and Huron waters, as Bretz indicated, must have escaped eastward. Because most of the Great Lakes region was reglaciated during the succeeding Mankato advance, the path of escape may never be

[2] A local topographic departure exists in the narrow sector between Binghamton and Sidney, New York, where the border mapped by MacClintock and Apfel lies some 12 miles northwest of the line which topography suggests might be expected.

known. That a choice of low-level routes across Ontario existed, however, seems almost beyond doubt. Furthermore, whether it carried the discharge from the upper lakes or only its own water, the lake in the Erie basin escaped eastward, for it could not have drained to the west. The lowest path of escape led over the Niagara escarpment into the basin of Lake Ontario. I suggest that the lake that occupied the Erie basin during the Two Creeks interval followed this path and established the early Niagara Falls represented today by the St. David drift-filled gorge (map in Kindle and Taylor, 1913, p. 21), the head of which was reoccupied and cleared of its drift filling by the Niagara River of post-Mankato-maximum time.

Taylor ascribed the St. David gorge to the recession of an interglacial Niagara Falls which suddenly ceased as a result of glacial advance; this view was confirmed with slight modification by Johnston (1928), who considered the gorge pre-Wisconsin and apparently did not entertain the alternative possibility of an intra-Wisconsin date. I have found nothing in the logs of borings published by him, nor in his discussion, that precludes the interpretation that the St. David gorge was cut during the Two Creeks interval and then was packed full of drift by the succeeding Mankato advance.

Probable St. Lawrence Marine Submergence

A glacier margin north of the upper St. Lawrence in Two Creeks time has been shown to be probable. The position of the glacier margin with respect to the lower St. Lawrence at that time is not determinable from any evidence known to me at present. However, both the isobases on the water surfaces of the later marine invasions (Flint, 1947, p. 420) and the Cary(?) drift border in New England have a parallel, northeast trend, suggesting deglaciation toward the northwest rather than toward the north. Measured from the Cary(?) maximum in New England, a retreat of the glacier margin northwestward amounting to about 300 miles should have been sufficient to bring the sea into the St. Lawrence lowland, presumably still deeply depressed as a result of glacial loading.

Any marine deposits made at that time, and not later removed by erosion, must lie buried beneath Mankato drift. Along the St. Lawrence, several occurrences of fossil-bearing marine sediments plowed up and overlain by drift were reported by Coleman (1927; 1932). Those in the upper St. Lawrence region, as at Cornwall, Ontario, and Waddington, New York, cannot be related to the Champlain Sea or the Ottawa Sea because at this longitude those seas created shore features that were not later overridden by the ice sheet. It is, furthermore, improbable that the array of marine deposits recorded by Coleman could have survived Sangamon interglacial erosion and repeated glaciations during the Wisconsin age. I suggest that the drift-covered marine sediments reported by Coleman in the upper St. Lawrence region (and possibly some of those along the lower St. Lawrence) are relicts of an early marine submergence, of Two Creeks date.

Aqueous Deposits at Toronto

The glacier margin of Two Creeks time, inferred in the preceding sections, would have stood at least 250 miles north of the vicinity of Toronto. Quite possibly the "latest interglacial" deposits (not the well-known "Toronto interglacial" beds) described by Coleman (1933, p. 2, 31–32) as occurring discontinuously from Toronto to Port Hope, Ontario, belong to the Two Creeks horizon. They consist of sand with *Amnicola limosa* (an aqueous gastropod) and *Succinea* (a terrestrial gastropod). Coleman stated that their occurrence implies climatic conditions much like those in the same district today, although actually this may mean little more than that the district was not immediately adjacent to the ice sheet. The sand lies between two tills, the upper one of which is the latest, or surface, till of the region. Coleman considered the sand interglacial, but cited no supporting evidence. On the basis of field study in 1947, Mr. A. K. Watt (personal communication) of the Ontario Department of Mines concluded that "there was no justification for any interval exceeding a sub-age for these sands," and stated that he knew of no obstacle to their being assigned to the Two Creeks horizon.

Probable Mankato Drift Border

Readvance in the Glens Falls-Fort Edward Area

The evidence indicates that at the close of the Two Creeks interval the glacier margin readvanced at least 275 miles in the Lake Michigan basin to a position south of Milwaukee, that it readvanced far into the Lake Huron basin, and that in the Lake Ontario basin it readvanced at least as far as Syracuse. In view of the evidence of general, substantial readvance it is improbable that the Mankato maximum position does not record readvance at the same time in eastern New York and New England. With readvance as a clue, it should be possible to narrow the search for the Mankato drift border in the eastern region.

The published descriptions of the varved silt-and-clay body accumulated in Lake Albany in the Hudson valley report those sediments as undisturbed and as the latest glacial deposits in the valley (*e.g.*, Woodworth, 1905). The northern limit of the Lake Albany sediments is northeast of Glens Falls. Furthermore, northeast of Glens Falls lies the sequence of sediments and strandlines made first by Lake Vermont and later by the Champlain Sea, in the basin of Lake Champlain. These, too, are undisturbed and have not been overridden by glacial advance.

Therefore, throughout the 250-mile, north-south distance between the southern end of the Lake Albany deposits and the St. Lawrence River, the only sector in the Hudson-Champlain lowland where readvance can reasonably be looked for is the vicinity of Glens Falls. It is precisely in that vicinity, and only in that vicinity, where readvance has been reported. Woodworth (1905, p. 143–144, 151) noted that the lake-floor clay between Fort Ann and Moses Kill in the Glens Falls-Fort Edward area is plowed up or molded into knobs and hillocks by glacier ice and strewn with boulders. These effects extend through a maximum north-south distance of more than 15 miles. He stated (p. 144)

> "all the evidence points to the conclusion that after the deposition of clays over the Fort Edward district at a time when the ice front had retreated an unknown distance to the north, there was an advance of the ice probably as far south as the mouth of the Moses Kill within the Hudson gorge." (Moses Kill empties into the Hudson 5 miles south of Fort Edward.)

Confinement of the evidence of overriding to the area east of the Hudson is not surprising, as the area west of the Hudson is covered by the Glens Falls delta, built by the Adirondack Hudson into a later water body, Lake Verrmont (Chadwick, 1928).

Chapman (1937, p. 95) followed Woodworth's interpretation of Lake Albany and the succeeding water body, Lake Vermont, the earliest outlet of which, at Coveville, stood 15 miles south of Fort Edward. He did not, however, comment on the retreat and readvance that intervened, according to Woodworth's implication, between the Lake Albany and Lake Vermont episodes.

I suggest that the overriding is the record of the Mankato advance. The latitude of Glens Falls and Fort Edward is only 10 miles north of that of Syracuse, to which the Mankato ice is known to have advanced; the Champlain-Hudson lobe of an ice sheet that reached Syracuse west of the Adirondacks can reasonably be expected to have reached the vicinity of Glens Falls on the eastern side of that mountain mass. If the Mankato border lies a short distance south of Fort Edward, the advancing Champlain lobe of the Mankato glacier must have created a lake in the Champlain basin—an early, rising phase of Lake Vermont, hitherto unrecognized.

Readvance in the St. Johnsbury and Gorham Districts

Evidence of readvance in the St. Johnsbury, Vermont, district and at the Junction of the Connecticut and Passumpsic rivers 8 miles south of St. Johnsbury is on record. Varved silt and clay in the St. Johnsbury district is disturbed and is interbedded with gravel. That fact coupled with the presence of "morainal deposits" led Antevs (1922, p. 84; 1928, p. 119, 120) to infer a readvance and prolonged stand of the glacier terminus in that district.

Observations of excavations connected with the Fifteen-Mile Falls Dam on the Connecticut River in Monroe, New Hampshire, just above the mouth of the Passumpsic River were made

by Crosby (1934) over a period of 5 years. He recorded, at numerous places on both sides of the river, sections exposing two till sheets separated by a maximum thickness of more than 30 feet of sand. The upper surface of the lower till had been channeled before deposition of the sand. Crosby inferred a readvance of the glacier margin following deglaciation of unknown amount, but not long enough to permit oxidation of the lower till before it was covered by the later sand.

Crosby (p. 419) further noted, on Peabody River a short distance southwest of Gorham, 25 miles east of Fifteen-Mile Falls, till overlying varved clay and silt, and he inferred a readvance of the glacier margin over proglacial lake deposits.

I suggest that the readvance features recorded near St. Johnsbury, at Fifteen-Mile Falls, and near Gorham are contemporaneous and that they represent the Mankato advance. It should be noted that the sequence of varves in the Connecticut Valley identified and numbered by Antevs was not carried northward through St. Johnsbury. The St. Johnsbury area marks the southern end of a conspicuous gap in the sequence (Antevs, 1928, p. 153); consequently the suggestion of a major retreat and a readvance to St. Johnsbury does not conflict with the record of the varves.

Evidence that no important readvance took place in the Connecticut Valley, in the segment Hartford-St. Johnsbury, is almost as clear as the similar negative evidence in the parallel Hudson Valley. The many sections of varved fine sediments, examined by a number of geologists through a period of 30 years, have brought to light no evidence of readvance other than a single section at Amherst, Massachusetts, where Antevs (1922, p. 18, 77–81, 97) reported 2½ feet of till overlying crumpled varved clay. In their glacial mapping of the Connecticut Valley in Massachusetts, Jahns and Willard (1942) did not mention either this section or any similar features; presumably therefore the section observed by Antevs is a phenomenon not duplicated in other exposures in the vicinity.

Pollen Profiles from Aroostook County

Pollen profiles reconstructed from stratigraphic sections of the sediments of three lakes and one bog in the northern part of Aroostook County, Maine, reveal a sequence of climatic changes strikingly similar to those implied by the pollen profiles from continental Scandinavia (Deevey, 1951). They exhibit a tundra zone, a slightly warmer zone of Alleröd character, and zones recording a postglacial thermal maximum followed by a recurrence of colder climate. The Alleröd horizon of northern Europe is known through radiocarbon dating to be the equivalent of the Two Creeks horizon in North America (Flint and Deevey, 1951). In the Aroostook County pollen sections, this horizon lies near the base of the section, separated from the underlying glacial drift only by the tundra zone of the pollen sequence. This being the case, the underlying drift (the surface drift of the district) is probably Cary, and therefore the outer limit of the Mankato drift lies northwest of the Aroostook County pollen localities. These relationships are a good example of the value of reliable pollen studies to the stratigraphy of glacial drifts.

Suggested Correlation

In view of the evidence of readvance in the Glens Falls-Fort Edward district, the St. Johnsbury-Fifteen-Mile Falls district, and the vicinity of Gorham, and in view of the pollen evidence from Aroostook County, I suggest that the Mankato drift border passes through the first three districts and lies somewhere between the Aroostook County pollen localities and the St. Lawrence River.

The line suggested has a northeast trend and possesses two features, independent of those described above, that make it seem reasonable: (1) It is very nearly at right angles to the striations plotted on the Glacial Map of North America, and (2) it parallels generally the marine isobases of the region (Flint, 1947, p. 420).

The position of the suggested line, not far southeast of the St. Lawrence, lends probability to the view that deglaciation during the Two Creeks interval was sufficient to flood the St. Lawrence Valley with sea water. If the Mankato readvance to the position suggested was only half as great as the minimum known Mankato readvance in the Lake Michigan basin—that is, half of 275 miles—it must have started from a line well north of the St. Lawrence, for the St.

Lawrence is only 90 miles northwest of northern Aroostook County.

Much field study will be needed to fix the position of the Mankato drift border between the localities mentioned here. The problem is complicated by the probability that local centers of glacial outflow, some of them connected with the ice sheet and some of them independent of it, occupied the Adirondacks and a number of highlands in northern New England, the Maritime Provinces, and Quebec (Flint, 1951a). With the aid of pollen profiles made at critical localities, however, it should be possible to fill in the limit of the region covered by Mankato glacier ice.

The faintness of the Mankato drift border in New England probably is attributable to the same factors—high relief, steep slopes, resistant rocks, and scanty drift—that made the Cary border inconspicuous. In fact the Mankato border seems even fainter than the Cary; this difference may be the consequence of thinning of the ice sheet between the Cary and Mankato maxima. The effect would have been to increase relief relative to ice thickness and to restrict still further the freedom of flow of the glacier, the rate of erosion of the bedrock, and the bulk of ice-transported debris available for building moraines and outwash.

The ice caps suggested on the map (Pl. 2) as late Cary(?) are correlated with that substage on the following basis: The latest ice movement in southeastern Quebec was southeastward and was preceded by a movement toward the northwest. The latest ice to cover this region is believed to have been the Mankato Laurentide Ice Sheet; the earlier, local flow may record highland ice caps formed during the Mankato advance. (For fuller discussion, *see* Flint, 1951a.)

Inferred Events from the Mankato Maximum to the Present Time

Correlation Chart

Events since the Mankato maximum have consisted principally of: (1) glacial readvances, (2) successive lake phases, (3) marine invasions of the St. Lawrence lowland, (4) successive phases in the recession of Niagara Falls, and (5) alternating phases of crustal quiescence and uplift as deglaciation proceeded irregularly. These events are brought together in a chart (Pl. 3) in order to avoid tedious chronologic presentation in the text. The discussion that follows, therefore, is not chronologic, but deals rather with the mutual relations of the various kinds of evidence. Throughout the discussion it is assumed that the reader has the chart before him.

Marine Phases in the St. Lawrence Lowland

General statement.—The stratigraphy of the Pleistocene marine deposits in the St. Lawrence lowland is represented by a long and complicated literature extending back into the early nineteenth century. The present discussion merely summarizes the present rather unsatisfactory position of stratigraphic correlation.

Until very recently the marine sequence was subdivided into a lower member, the "Leda clay" (including much silt), and an upper member, the "Saxicava sand." This subdivision has proved to be misleading, because the pelecypods *Leda* (*Nuculana*) and *Saxicava* apparently occur in both members. Furthermore it is probable that the distinction between clay and sand is not everywhere stratigraphic, but that each marine invasion is represented by both a clay facies and a coarser (littoral?) facies.

It is evident that at least two distinct bodies of marine sediments, separated by an unconformity, occupy the St. Lawrence lowland. These are, respectively, the Champlain marine deposits and the Ottawa marine deposits. The older body, a record of the Champlain Sea, postdates at least part of the history of Lake Algonquin.

Champlain Sea.—The Champlain Sea succeeded Lake Vermont in the Champlain basin (Chapman, 1937). It flooded the Lake Ontario basin and submerged the eastern end of the Trent Valley outlet of Lake Algonquin, after the extinction of Lake Iroquois (Johnston, 1917, p. 32). Present-day depth range of the invertebrates common in the basal part of the marine deposits, compared with the range of altitude through which the deposits occur, implies that the marine water rose against the land—this must have been a eustatic movement resulting from rapid glacial melting (Johnston, 1917, p. 30–31)—and reached a point represented by the present altitude of 690 feet near Ottawa. During this time the

crust remained quiescent or rose only very slowly. In places, therefore, the sea water was more than 500 feet deep. The lower part of the Champlain deposits contains many arctic forms; the upper part is nearly barren of fossils, but those present imply less cold or less deep water. Furthermore the lower part incloses many erratic boulders and records much disturbance through shove by grounding icebergs; the upper part is finer-grained, shows no disturbance, and contains few erratics. These observations suggest that during the marine invasion the glacier was receding into the territory north of the St. Lawrence.

Ottawa Sea.—The Ottawa Sea followed the Champlain Sea, but in the upper St. Lawrence region, at least, emergence accompanied by erosion intervened between the two marine invasions. How much of the St. Lawrence Valley was affected by the emergence is not known, although there is evidence that emergence occurred at least as far down the valley as Quebec City.

Antevs and Goldthwait (Antevs, 1939, p. 712) showed that two distinct bodies of marine sediment, separated by an unconformity, are exposed at a number of localities in the valley of the Ottawa River. At Ottawa, the older body reaches a present altitude of 690 feet, whereas the younger reaches to only 240 feet, at which altitude it is marked by a strand line. The younger sediments were ascribed by Antevs (1939) to an Ottawa Sea,[3] transgressing the land as a result of eustatic rise of the sea surface.

Fossils collected from marine deposits at low altitudes in the St. Lawrence lowland can be reasonably ascribed to the Ottawa Sea or to the upper part of the Champlain Sea, although the collections were made before the marine deposits were differentiated into the Champlain and Ottawa bodies and were described as from "Leda clay" or merely "marine beds." Johnston reported the collection of oysters (species not specified) from marine beds at Montreal (Coleman, 1932, p. 314).[4] Baker (1920, p. 178) reported many kinds of land plants, including sugar maple, balsam poplar, and yellow birch, and a number of freshwater plants, as well as mollusks and insects, from "Leda clay" from seven localities at low altitudes in and near Ottawa; the land-plant assemblage was stated to be similar to that living in the same area today. The oysters represent temperatures slightly higher than those prevailing today. Johnston (1917, pp. 25, 30) mentioned the occurrence in marine clay of chipmunk bones as well as leaves of plants, but as his study antedated the recognition of two distinct bodies of marine sediment, he suggested that such fossils might be intrusive into the marine deposits through recent slumping.

These fossils seem ecologically incompatible with the fauna of the lower part of the Champlain deposits. Inasmuch as they suggest a climate comparable with the present-day climate, they can be provisionally placed at or after the Thermal Maximum, a period dating roughly from 5000 years ago on European evidence, when inferred temperatures were slightly higher than those of today. Such placing implies that the Ottawa Sea, and perhaps also the last phase of the Champlain Sea, existed at or after the Thermal Maximum—admittedly on grounds that cannot be firm until more facts are known; the correlation is no more than a reasonable guess from existing data. However, both the upward change in the fauna of the Champlain sediments and the fossils implying temperate conditions fit reasonably into the concept of rise in temperature between the beginning of the Champlain marine invasion and the Ottawa invasion. The eustatic rise of sea level recorded by the Champlain deposits and, as noted by Antevs, in the Ottawa sediments expectably reflects such a temperature rise in terms of the rapid return of water substance from glaciers to the sea.

The relationships reported by Osborne (1951) from the Quebec City segment of the St. Lawrence Valley are rather similar to those evident in the Montreal-Ottawa region. There two bodies of marine deposits (p. 242) are separated by an unconformity apparently

[3] A possibly correlative marine invasion was termed by Osborne (1951, p. 243) the "Quebec Sea." That term is not used here because Woodworth (1905, p. 220) long ago showed it to be undesirable.

[4] The writer has been able to find no independent confirmation of this occurrence, and the possibility can not be dismissed that the shells are intrusive, having been deposited by Indians at a time postdating the Ottawa Sea. However, the land-plant assemblage alone adequately indicates the climate of Ottawa-Sea time.

caused by channeling by the St. Lawrence River, and a peat containing beetles (p. 237) may represent the intermarine episode of emergence and channeling. Possibly the two marine bodies are Champlain and Ottawa, respectively.

All these observations fit into the general scheme on the assumption that the Champlain sea breached the glacier ice remaining (perhaps as a local ice cap) in the lower St. Lawrence region, and flooded the valley while the region was still deeply depressed. Worldwide melting raised the sea level rapidly. Presently rapid upwarping began, raising the crust faster than the rate of rise of sea level. Emergence resulted, and subaerial erosion occurred. Crustal uplift then slowed, or ceased altogether, so that the continuing rise of the sea once more flooded the valley as the Ottawa Sea, this time much less deeply.

Low-water Phase in the Michigan and Huron Basins

According to the sequence of events set forth in the classic synthesis by Leverett and Taylor (1915), the Lake Algonquin phase passed into the Nipissing Great Lakes phase by direct transition. However, these authors stated (p. 447) that "Knowledge concerning the beginning of the Nipissing Great Lakes and the transition to them from Lake Algonquin is rather meager. . . ." In another place (p. 438) they said "It seems certain that an outlet for Lake Algonquin to the Ottawa Valley opened before the beginning of the Nipissing Great Lakes and the full eastward discharge over the col at North Bay, but the facts available at present are not decisive."

It was shown later by Stanley (1938) that a low-water phase occurred in the basins of Lakes Michigan and Huron after the late phases of Lake Algonquin but before the early phase of the Nipissing Great Lakes. The low-water phase, during which the water surfaces stood as much as 300 feet below their present positions, is recorded by a submerged valley through the Strait of Mackinac; its stratigraphic position is firmly established by the attitudes of the warped Algonquin and Nipissing strandlines.[5] Stanley (p. 972) inferred some undiscovered low eastward outlet or outlets in the region east of Georgian Bay.

A low-water phase at this time is expectable because, according to the radiocarbon dates, the Thermal Maximum had reached and passed its peak. The pronounced increase in warmth that led up to the Thermal Maximum, in this region the greatest recorded since the Two Creeks interval, had freed the St. Lawrence Valley of glacier ice sufficiently to bring the sea into that valley throughout its entire length. Deglaciation sufficient to free the St. Lawrence is likely to have freed every one of the known outlets of the glacial Great Lakes.

I suggest therefore that, at the end of Lake Algonquin time, deglaciation uncovered the North Bay-Ottawa River outlet that later became the outlet of the Nipissing Great Lakes. That outlet was then more, perhaps very much more, than 700 feet lower than it is now. The upper lakes were thereby drawn down to very low levels, the Strait of Mackinac was channeled by the stream draining Lake Michigan, and the whole discharge of three lakes was poured into the sea standing in the Ottawa River valley. Johnston (1917, p. 32) was puzzled by the enormous volume of the marine clay in the Ottawa region, sought for a source and concluded that the sediment must have been brought by the North Bay outlet-discharge of the Nipissing Great Lakes. However, the low-water phase in the upper lakes provides a more likely source. The phase of falling water levels would have exposed large areas of fine-grained lacustrine sediment to rainwash and gullying; the products of that erosion, carried in suspension and dumped into the sea, meet the requirement better than do the sediments of the more stable Nipissing lakes.

Johnston (p. 32) noted also that the marine clay extends far north up the valley of the Gatineau River, tributary to the Ottawa River at Ottawa, and is particularly thick opposite the mouth of the Gatineau. He inferred that the Gatineau had contributed importantly to the marine sediments. Had Johnston's study postdated the recognition of two marine invasions separated by an emergence, instead of only one,

[5] After this paper had been submitted for publication, additional evidence, drawn from sediments and paleontology, of this low-water phase was published by J. L. Hough (Geol. Soc. Am., Abstracts November 1952 meeting, p. 53).

undoubtedly he would have considered the possibility that the sediments were carried up the Gatineau during the Champlain submergence and, during the following emergence, were progressively laid bare, reworked, carried down the Gatineau, and thickly redeposited near Ottawa.

Nipissing Great Lakes

While the Ottawa Sea was invading the St. Lawrence region, continuing crustal uplift gradually raised the North Bay outlet channel, causing the levels of the upper lakes to rise with it and introducing the Nipissing phase. With continued upwarping, these lakes rose against their southern shores, the Detroit outlet was re-established, and for a brief time both outlets functioned. Then the North Bay outlet, still rising, dried up, and with that event the modern Great Lakes came into existence. The same uplift caused the regression of the Ottawa Sea.

Lake Agassiz

Glacial Lake Agassiz came into existence when the retreating margin of the Red River lobe of the Mankato ice sheet uncovered the vicinity of Lat. 46° N., Long. 96° W., in extreme western Minnesota. Possibly the lake formed at about the same time as Lake Duluth in the Lake Superior basin. Like the glacial Great Lakes, Lake Agassiz enlarged northward as the glacier receded, and during the earlier part of its history it discharged southward into the Mississippi River system.

During the earlier systematic study of Lake Agassiz it was supposed that the lake enlarged to its greatest extent (the Hayes River-Nelson River region on the slope leading to the southwest shore of James Bay), becoming lower by stages as new, eastward outlets were uncovered, and then came to a permanent end as the uncovering of a very low outlet drained away its water.

Later, however, Johnston (*e.g.* 1946) held that a widespread unconformity within the lake-floor sediments and a break in the sequence of the beach-ridge strandlines demonstrate draining and refilling of the lake and reactivation of its southern outlet. An earlier and shorter desiccation of the lake was demonstrated by the discovery of a "drying surface" lower in the stratigraphic section of lacustrine deposits (Rominger and Rutledge, 1952, p. 176). It is generally agreed that the two episodes of desiccation were brought about by recession of the glacier that constituted the northeastern shore of the lake, the temporary uncovering of low eastward outlets, (probably across southwestern Ontario to Lake Superior), and the readvance and closing of those outlets.

The evidence of the lower "drying surface" indicates the first uncovering and readvance as a temporary affair. The later readvance, however, involved a movement of greater magnitude. Johnston (p. 11) thought it probable, through the mutual relations of the unconformity, a certain strand line, and an end moraine, that the second readvance brought the margin of the ice sheet to the Pas moraine west of Lake Winnipeg. Again through shoreline/moraine relationships, he thought it likely that the end moraines near Dryden and Ignace, north of Lake Superior, were built at the limit of the same readvance. The correlation is reasonable in that the implied drift border trends at right angles to the striations in the region it traverses. Also the implied border parallels the isobases on the Lake Agassiz strand lines and parallels the general trend (although not the individual lobations) of the Mankato drift border. But the correlation has not been established, nor has the position of the ice margin in eastern Ontario and Quebec, at the time of this readvance, been determined.

Having in mind this possible outer limit of readvance in Manitoba, we may note the evidence in the Thelon River region, west of Hudson Bay, of a west-northwest movement of the ice sheet, followed by a glacial lake apparently dammed by ice on the east, followed in turn by a north-northwest movement, all within "the closing stages of the Ice Age" (Bird, 1951). This assemblage invites the speculation that the earlier striations represent the retreat during the time of the earlier Lake Agassiz, that the lake was a correlative of the earlier Agassiz and was formed in a similar manner, and that the later striations, which reach only as far west as the 99th Meridian, were made by the readvance identified by Johnston with the Pas moraine.

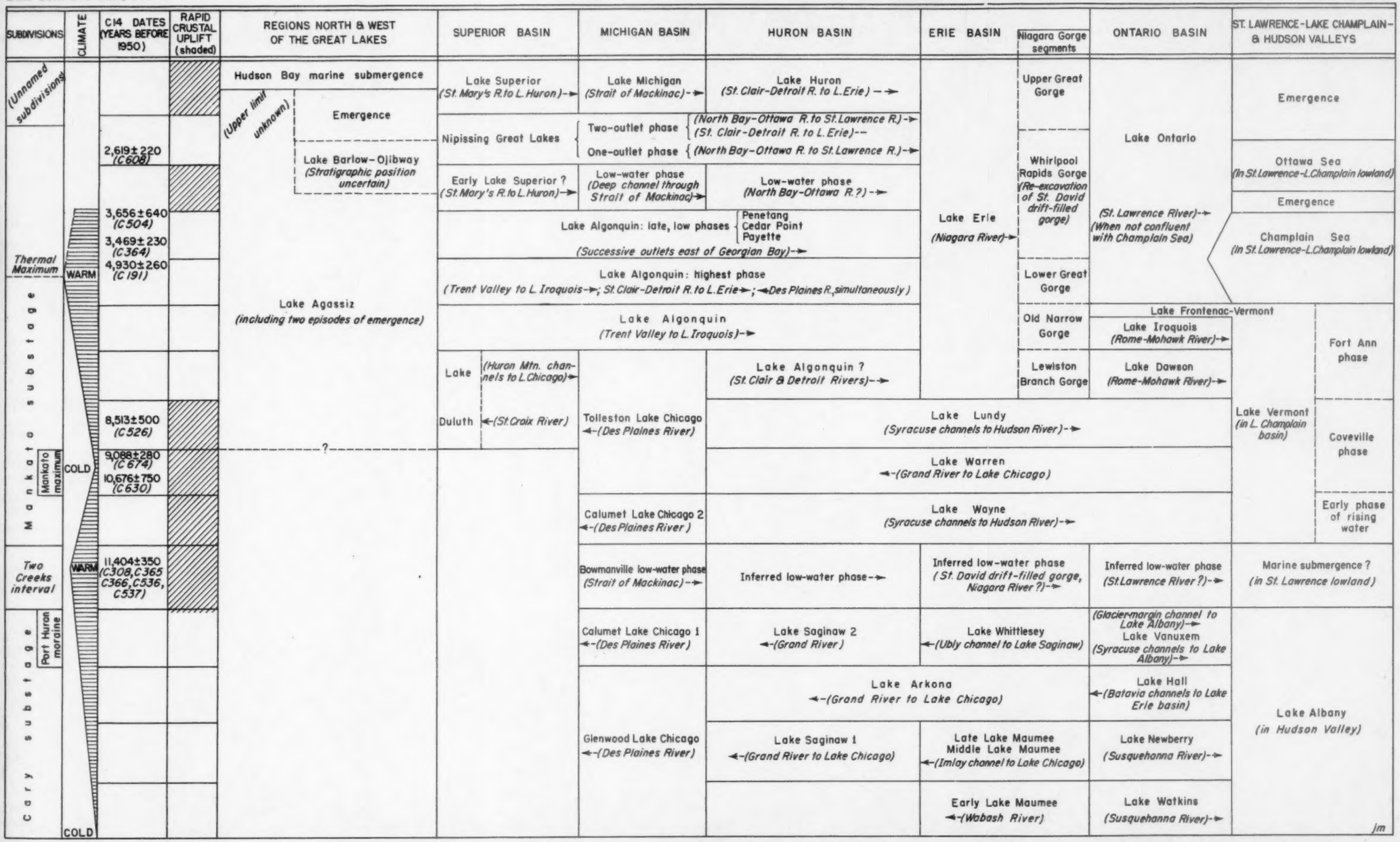
SUBDIVISIONS
CLIMATE
C14 DATES (YEARS BEFORE 1950)
RAPID CRUSTAL UPLIFT (shaded)
REGIONS NORTH & WEST OF THE GREAT LAKES
SUPERIOR BASIN
MICHIGAN BASIN
HURON BASIN
ERIE BASIN
Niagara Gorge segments
ONTARIO BASIN
ST. LAWRENCE-LAKE CHAMPLAIN-& HUDSON VALLEYS
(Unnamed subdivisions)
Thermal Maximum
Mankato substage
Mankato maximum
Two Creeks interval
Cary substage
Port Huron moraine
WARM
COLD
WARM
COLD
2,619±220 (C608)
3,656±640 (C504)
3,469±230 (C364)
4,930±260 (C 191)
8,513±500 (C526)
9,088±280 (C 674)
10,676±750 (C630)
11,404±350 (C308, C365 C366, C536, C537)
Hudson Bay marine submergence
(Upper limit unknown)
Emergence
Lake Barlow-Ojibway (Stratigraphic position uncertain)
Lake Agassiz (including two episodes of emergence)
?
Lake Superior (St. Mary's R. to L. Huron)
Nipissing Great Lakes
Two-outlet phase
One-outlet phase
Early Lake Superior ? (St. Mary's R. to L. Huron)
Lake Algonquin: late, low phases
Penetang
Cedar Point
Payette
(Successive outlets east of Georgian Bay)
Lake Algonquin: highest phase
(Trent Valley to L. Iroquois; St. Clair-Detroit R. to L. Erie; Des Plaines R., simultaneously)
Lake Algonquin (Trent Valley to L. Iroquois)
Lake
Duluth
(Huron Mtn. channels to L. Chicago)
(St. Croix River)
Lake Michigan (Strait of Mackinac)
(North Bay-Ottawa R. to St. Lawrence R.)
(St. Clair-Detroit R. to L. Erie)
(North Bay-Ottawa R. to St. Lawrence R.)
Low-water phase (Deep channel through Strait of Mackinac)
Tolleston Lake Chicago (Des Plaines River)
Calumet Lake Chicago 2 (Des Plaines River)
Bowmanville low-water phase (Strait of Mackinac)
Calumet Lake Chicago 1 (Des Plaines River)
Glenwood Lake Chicago (Des Plaines River)
Lake Huron (St. Clair-Detroit R. to L. Erie)
Low-water phase (North Bay-Ottawa R.?)
Lake Algonquin ? (St. Clair & Detroit Rivers)
Lake Lundy (Syracuse channels to Hudson River)
Lake Warren (Grand River to Lake Chicago)
Lake Wayne (Syracuse channels to Hudson River)
Inferred low-water phase
Lake Saginaw 2 (Grand River)
Lake Arkona (Grand River to Lake Chicago)
Lake Saginaw 1 (Grand River to Lake Chicago)
Lake Erie (Niagara River)
Inferred low-water phase (St. David drift-filled gorge, Niagara River ?)
Lake Whittlesey (Ubly channel to Lake Saginaw)
Late Lake Maumee
Middle Lake Maumee (Imlay channel to Lake Chicago)
Early Lake Maumee (Wabash River)
Upper Great Gorge
Whirlpool Rapids Gorge (Re-excavation of St. David drift-filled gorge)
Lower Great Gorge
Old Narrow Gorge
Lewiston Branch Gorge
Lake Ontario
(St. Lawrence River)
(When not confluent with Champlain Sea)
Lake Frontenac-Vermont
Lake Iroquois (Rome-Mohawk River)
Lake Dawson (Rome-Mohawk River)
Inferred low-water phase (St. Lawrence River ?)
(Glacier-margin channel to Lake Albany)
Lake Vanuxem (Syracuse channels to Lake Albany)
Lake Hall (Batavia channels to Lake Erie basin)
Lake Newberry (Susquehanna River)
Lake Watkins (Susquehanna River)
Emergence
Ottawa Sea (In St. Lawrence-L. Champlain lowland)
Emergence
Champlain Sea (In St. Lawrence-L. Champlain lowland)
Lake Vermont (in L. Champlain basin)
Fort Ann phase
Coveville phase
Early phase of rising water
Marine submergence ? (in St. Lawrence lowland)
Lake Albany (in Hudson Valley)
jm

CHART SHOWING SEQUENCE, MUTUAL RELATIONS, AND OUTLETS OF THE GREAT LAKES, AND FEATURES RELATED TO THEM

Inferred from the relations set forth in the text Undoubtedly the actual sequence was more complicated.

R–
M–
R. V.
C–

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FIGURE 1.—SKETCH MAP SHOWING RELATIONS OF LAKES AGASSIZ AND BARLOW-OJIBWAY
R—Possible limit of late readvance in Lake Agassiz region.
M—Mankato drift border (approximate; modified after Glacial Map of North America; in Iowa after R. V. Ruhe).
C—Cary drift border (approximate; modified after Glacial Map of North America).

Terrain between Great Lakes and James Bay

Very little is known about the deglaciation of the terrain north of the Great Lakes region. When the Champlain Sea first invaded the Ottawa district, the ice-sheet margin was in the vicinity and calved into the water, as the evidence of erratics and grounding icebergs attests. Later in Champlain time, the glacier had shrunk well to the north, for these features are lacking in the topmost Champlain deposits.

During the retreat, the ice created the vast Lake Barlow-Ojibway (Antevs, 1925, p. 75), well north of the Great Lakes. The lake could have existed only while the glacier constituted its northern shore, as the country slopes north to James Bay. In it varved silt and clay accumulated. Further deglaciation then caused complete drainage of the lake (Antevs, 1928, p. 103). How far north of the Barlow-Ojibway lake deposits the glacier margin stood when at its maximum position is unknown.

The literature indicates that after the lake had been drained the glacier margin readvanced, depositing till in places on the varved fine sediments of the former lake. Till overlying lacustrine sediments has been reported by Antevs (1928, p. 104; 1931, p. 19), Hills (1947), Tanton (1919, p. 49), and Imbault (1952, p. 63).[6] Antevs inferred a readvance at least to a position 33 miles south of Cochrane, Ontario, and a total minimum readvance of 70 miles, with the maximum value unknown.

The relation of this readvance, inferred by Antevs, to the sequence of events in the Lake Agassiz basin is not yet known.

In this connection we may note the peat and inorganic sediments occurring between two tills, one of which is the surface till, at many localities in the region southwest of James Bay. Plants identified from the peat include spruce, pine, birch, and fir, willow, sedges, and mosses—an assemblage much like that living in the same region today. Originally the peat was considered interglacial (reference in Flint, 1947, p. 287), but it may date from late in the Wisconsin age, the overlying till having been deposited by a very late readvance of the glacier margin.

Similarly the marine clay overlain by till at Moose Factory and along the Albany River, in the same general region, may be a further record of a late deglaciation and a succeeding readvance, as suggested by Flint (1947, p. 288).

Evidence assembled by Osborne (1951) indicates that very late in the process of deglaciation a local ice cap flowed out from the highlands north of Quebec City and deposited till over the Champlain marine sediments. Osborne suggested (p. 242) that the ice cap may correlate with the readvance in the vicinity of Cochrane. However, as no firm connection between southern Quebec and the region south of James Bay has been established, events in the three regions—Manitoba, eastern Ontario, and southern Quebec—remain uncorrelated.

In the region south of James Bay, striations and eskers (Flint and others, 1945) suggest that the latest recorded flow of ice was southward from the direction of James Bay. This movement implies that at that time the Laurentide Ice Sheet, as such, still existed. However, the trends of striations and eskers farther east suggest the possibility that, after ice in the Hudson Bay region had disappeared, active ice caps persisted on the highlands of eastern Quebec and Labrador. Apparently the "Quebec monadnock" ice cap inferred by Osborne (1951) was one of these. The present-day climate of this highland, as described by Osborne (p. 229), and of this and other highlands as described by Leighly (1949) supports the concept of a very late persistence of glaciers in this region (Flint, 1951a). It may be suggested reasonably that the ice cap radiating from the Gaspe highlands dates from the Mankato maximum, and that the later cirque glaciers in those highlands record a later re-formation of glacier ice.

Sequence of Climatic Fluctuations

The sequence of climatic fluctuations in northeastern America during the Wisconsin glacial age is recorded by (1) end moraines, overridden lake deposits, low-water lacustrine phases, proximal limits of lake strandlines, and similar evidence of fluctuation of the ice-sheet margin; (2) chemical decomposition of drift; (3) pollen profiles; (4) fossils in marine deposits; and (5) frost-disturbed mantle, ventifacts, and other "periglacial" features.

In the southern New England-Long Island region, the greater prevalence of periglacial phenomena in the outer belt of drift (Iowan-Tazewell complex?) than in the inner (Cary?) belt testifies to a cold, windy climate at the time when the Laurentide Ice Sheet was thickest and most extensive, and to less rigorous conditions when, after retreat and readvance to the Cary(?) maximum, the glacier was thinner and somewhat less extensive. The lobation of the Cary(?) drift border, inferred from the scanty data available, accords well with the strong lobation of the Cary border, relative to the Iowan and Tazewell borders, in and west of the Great Lakes region.

The weathering alteration (chiefly oxidation) of the upper part of the drift on which the Cary(?) drift lies, noted at a number of localities, records an interval of deglaciation more

[6] Since completion of this paper I have been informed by F. F. Osborne that till overlying Barlow-Ojibway deposits has recently been reported also in the Allard River area, Quebec. J. A. Elson (personal communication) questions whether the field data in the Cochrane area actually record readvance. The question remains unsettled.

conspicuous than any other so far identified within the Wisconsin stage. It seems likely—although this is pure speculation—that, during this (Brady) interval, the ice sheet shrank back north of the Great Lakes. The subsequent Cary(?) readvance marks a pronounced low in the regional temperature record, and the Two Creeks interval, recorded by a low-water phase and an inter-till peat layer, marks another high, possibly of shorter duration than the Brady high because no weathering at the Two Creeks horizon has come to light, and because continuity of sedimentation at the type locality is indicated (Flint and Deevey, 1951, p. 262).

Fossils collected from the marine sediments in the St. Lawrence lowland suggest, from Champlain time to Ottawa time, a sea-temperature rise that may have lagged considerably behind the rise in atmospheric temperatures.

Although the only pollen profiles that specifically support the inferred sequence of temperatures beginning with the retreat from the Cary maximum are those from northern Aroostook County, Maine (Deevey, 1951), pollen profiles from elsewhere in New England are not in conflict with it. The lakes and bogs from which the profiles were reconstructed preserve a record of continuous sedimentation, which, in the case of the localities described from the New Haven region (Deevey, 1939), should have begun after the Tazewell maximum. The rise of spruce pollen, by means of which Zone A2 was defined in those profiles, was interpreted by Deevey as a record of cooling climate attending a glacial readvance to a position near New Haven; that interpretation would agree with the position of the Cary(?) substage as inferred in this paper. However, as a result of the discovery of "Late-glacial" (tundra) zones in the Aroostook County localities, Deevey (1951, p. 202–203) concluded that the spruce-pollen maximum in the southern Connecticut profiles is probably only the statistical consequence of ignoring non-arboreal pollen, and that Zone A2 records the first genuine appearance of spruce forest in the region after a tundra phase when appreciable amounts of spruce pollen were wind-blown from a considerable distance.

More direct evidence of tundra-pollen zones from southern New England profiles is still lacking, except for a later boring in Upper Linsley Pond (Deevey and Potzger, 1951, p. 493). In this boring non-arboreal pollen not only is present in amounts characteristic of "Late-glacial" zones in Europe and in Aroostook County, but rises to a maximum before it falls, suggesting a return to glacial conditions that agrees with the concept of the Cary(?) maximum adopted here.

A complete re-examination of New England pollen sequences is required, and it seems likely that strategically located borings will provide the record of climatic fluctuations since the Tazewell maximum. As pollen diagrams from localities on Tazewell(?) drift in southern Connecticut give no hint of cooling during the *Mankato*(?) readvance, it is not to be expected that all profiles on Cary(?) drift will do so; in other words, the available evidence in pollen profiles from southern and central New England is not relevant to the scheme inferred here, partly because the older pollen zones are inadequately studied and partly because the borings were made in places too far outside the Cary(?) and Mankato(?) drift borders to give a complete record of the related climatic fluctuations.[7]

Sequence of Crustal Uplifts

It has long been known that the movements of upwarping in the Great Lakes region were pulsatory, and that the pulses of rapid uplift were separated by times of slower warping. Leverett and Taylor (1915, p. 506) identified three pulses of uplift, dated as follows by the attitudes of the warped strandlines: (1) Post-Whittlesey, pre-Algonquin, (2) Post-Algonquin, pre-Nipissing (the greatest of the three), (3) Post-Nipissing (the smallest of the three). The same authors (p. 518) expressed doubt as to the extent to which deglaciation was responsible for the three crustal movements, and added that "the difficulties in the way of a complete demonstration of this as the main cause seem well-nigh insuperable."

In Plate 3 the temperature factor in the climate, as inferred from glaciation and deglaciation, fossils, pollen profiles, and the like, can be compared with crustal uplift in the

[7] The three paragraphs ending with this footnote were contributed by E. S. Deevey, Jr.

time sequence. It will be seen that the first two pulses of rapid uplift began during times of warmth, when melting and thinning of the Laurentide Ice Sheet, and consequently the rate of unloading of the crust, must have been more rapid than at other times. This correlation supports the concept that glacial unloading of the crust was an important factor in the observed deformations. It would not be justifiable to assert that deglaciation was the sole cause, but it looks as though it were perhaps the most important cause.

Comparison of uplifts with glacial regimens shows further that the earliest of the three rapid responses of the crust to unloading lagged behind the unloading by an estimated 2000 years. That uplift, which commenced with the retreat from the Port Huron moraine, a late-Cary event leading to the Two Creeks recession, quite evidently was not halted by the Mankato readvance, but continued through the close of the Lake Lundy phase, by which time the ice sheet had shrunk back considerably from its Mankato-maximum position. Then, while post-Lundy deglaciation continued, rapid uplift ended, and the crust rose only slowly until the end of Algonquin time. We may speculate that the cause of the reduction in rate of uplift was the new loading caused by the Mankato advance—by the addition to the body of the glacier of a belt of ice more than 300 miles wide and more than 2000 feet thick, plus the accumulation of a considerable volume of water in the form of proglacial lakes—but that the resulting stress had to accumulate for some 2000 years before it succeeded in reducing importantly the rate of uplift.

The second or post-Algonquin pulse of rapid uplift could not have ceased at the beginning of Nipissing time, because the shift from the early, One-outlet phase to the later, Two-outlet phase of the Nipissing lakes was caused by gradual upwarping. Hence the break shown in Plate 3 between the second and third movements of rapid uplift seems to represent a second retardation of a continuous movement. Possibly such retardation was brought about by readvance of the glacier margin, such as that which postdated Lake Barlow-Ojibway.

Leverett and Taylor (p. 506) allowed for the possibility of pre-Whittlesey uplifts but could find no record of them. There is indeed no record in the Great Lakes region, because the Cary readvance that followed the Brady interval would have destroyed any morphologic evidence of pre-Cary Great Lakes; only the stratigraphic evidence—and that probably fragmentary—could have remained. However, if the correlation suggested in the present paper is accepted, there is evidence of pre-Cary upwarping along the New England coast, in the form of marine deposits uplifted and later overridden by the Cary readvance.

It is tempting to correlate the three former sea-level surfaces in Newfoundland, now warped up toward the northwest (Flint, 1940), with the three groups of warped strand lines in the Great Lakes region. However, no single correlation appears obvious, and the progress of deglaciation of the two regions may have been sufficiently different to preclude synchrony of events. We do not imply that the warm intervals did not affect Newfoundland, but we cannot be sure yet that thick ice still covered Newfoundland during one or both of the last two intervals of warmth.

Record of the Niagara Gorge

The relation between the varying discharge of Niagara River, the consequent width of the gorge left by recession of the falls, and the history of the tributary lakes was first noted in 1913 by Taylor, who later (1929, p. 258–261) revised the data and assembled them in a table. This was further revised and discussed by Antevs (1931, p. 20–24). The revised table, with some further modification, is included in the chart, Plate 3.

The cutting of the Upper Great Gorge has brought forth much discussion. That gorge segment had its inception at the beginning of the Two-outlet phase of the Nipissing lakes, when the discharge of the three upper lakes began to join that of Lake Erie and swelled the volume of the Niagara River. Soon afterward uplift closed the North Bay outlet, diverting the entire discharge to the Niagara route; hence cutting of the Upper Great Gorge continues thr h the present time. Taylor (Kindle and Taylor, '3, p. 24) estimated the age of the segment at 3000 to 3500 years, based on the current rate of recession of the Horseshoe Falls. Despite a later revision by Boyd (Johns-

ton, 1928, p. 29), which raised the figure to 4000 years or more, Taylor's estimate seems to have been the better of the two. On the basis of the radiocarbon date of material believed to date from One-outlet Nipissing (Dreimanis, 1952, p. 73), the cutting of the Upper Great Gorge should have begun less than about 2600 years ago. On the basis that sample No. 504 (mentioned in the section following) is no younger than the One-outlet Nipissing phase, cutting should have begun less than about 3500 years ago.

Radiocarbon Dates

The radiocarbon dates pertinent to the post-Cary history of the Great Lakes are shown in their approximate stratigraphic positions in Plate 3, together with their accession numbers (Libby, 1952; W. F. Libby, unpublished data) No. 608 has been discussed by Dreimanis (1952) who assigned it to One-outlet Nipissing. Nos. 191, 504, and 364 have been briefly discussed by Flint and Deevey (1951, p. 283–285). No. 191 was placed by them as not younger than the highest phase of Lake Algonquin. They followed Taylor in assigning No. 504 to One-outlet Nipissing and, noting that No. 364 could be either Two-outlet Nipissing or highest Algonquin, they suggested the former as the more probable because of the age agreement with No. 504. However, recent study by L. J. Chapman (personal communication) suggests that the peat layer from which No. 504 was collected is more likely very late Algonquin than early Nipissing. If this correlation should be established, then No. 364 is probably also Algonquin, a possibility recognized in the original discussion. The two samples are placed provisionally in this position, in Table 3.

No. 526 is a piece of wood from a fossil forest submerged apparently by the encroachment of early Lake Lundy, which deposited fine sediment over the forest horizon (R. P. Goldthwait, unpublished data).

No. 630 is a piece of wood taken from varved fine sediment at Kimberley, Wisconsin, and believed to be driftwood in Lake Oshkosh, a temporary lake dammed against the terminus of the Mankato ice sheet very early in its retreat. Hence this sample should measure an event that occurred almost immediately after the Mankato maximum in Wisconsin.

No. 674 is a piece of driftwood from Lake Chicago sediments identified as probably deposited during the Tolleston phase of the lake (J H. Bretz and Leland Horberg, personal communication). According to the correlation established by Bretz (1951b), this date should measure a time slightly later than the Mankato maximum.

The eight dates are consistent internally, and the Nipissing date agrees fairly well with Taylor's estimate of the age of the Upper Great Gorge of the Niagara River. As a group they seem to be satisfactory.

Conclusion

The suggested drift borders shown in Plates 1 and 2, and the tentative correlation scheme in Plate 3 should be accorded no more value than they pretend to have, namely, reasonable syntheses derived mainly from published reports, and having apparent internal consistency. Like all such syntheses, these may well cease to be reasonable when more facts become known. The later history of the St. Lawrence valley and the region north of the Great Lakes are so little understood at present that in these areas great future improvements in our knowledge can be expected. Perhaps the array of data brought together in this paper will add stimulus to the search for an improved synthesis.

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MINERALOGY AND PETROLOGY OF SOME ORDOVICIAN K-BENTONITES AND RELATED LIMESTONES

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ABSTRACT

Twenty-two samples of Ordovician K-bentonite were examined by the use of size analyses, heavy- and light-mineral separates, thin sections, differential thermal curves, x-ray diffraction patterns, chemical analyses, and electron micrographs. The less than one micron fraction of two typical samples was studied in detail. The K-bentonite consists of randomly interstratified expanded and nonexpanded 2:1 layers in the ratio of 1:4. In addition, many of the samples contain packets of chlorite, the cause of an endothermal peak on the differential thermal curves formerly believed due to illite. Chemical analyses of the two samples indicate a direct correlation between the percentage of K present and the percentage of nonexpanded layers (illite).

The nature of the heavy and light minerals shows that the clay altered from a volcanic glass. It is believed the clay formed as an expanded 2:1 mineral (montmorillonite) and later adsorbed K which caused 80 per cent of the layers to become nonexpanded (illite).

Twenty-seven samples of the limestone on either side of a series of K-bentonite beds were examined in detail. The heavy and light minerals indicate that the insoluble residue is composed of both volcanic and nonvolcanic material. The nonvolcanic material is largely a nonexpanded dioctahedral 2:1 clay (illite); the volcanic material altered to form chlorite. The chlorite and volcanic heavy minerals increase in abundance as the K-bentonites are approached. It is believed that the ash altered to chlorite because of the availability of Mg, which was probably enhanced by the presence of the calcite ooze.

The "Tioga Bentonite", a thin clay bed in the Middle Devonian, is nearly identical to the Ordovician K-bentonite.

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ORDOVICIAN K-BENTONITE

Introduction

Acknowledgment.—The writer is indebted to Thomas F. Bates for giving freely of his time and ability in the preparation of this paper. Thanks are due Paul D. Krynine and John C. Griffiths for their suggestions and criticisms of various phases of this investigation. Several valuable suggestions and criticisms were received from W. F. Bradley.

The work was supported by a grant from the Geophysics Branch of The Office of Naval Research and is part of an investigation of the morphology, structure, and origin of fine-grained minerals.

Geological occurrence.—Clay formed from the alteration of glassy igneous material, usually a tuff or volcanic ash, is common in the Cretaceous and Tertiary rocks of Western United States where it is usually composed of montmorillonitic clay minerals. Nelson (1922, p. 606) first reported altered volcanic ash beds in the Eastern United States, in Kentucky, Alabama, Tennessee, and Virginia. Later Bonine and Honess (1929, p. 1), Rosenkrans (1933), and Kay (1931, p. 361; 1935, p. 225–244; 1944, p. 18–20) described occurrences in Central Pennsylvania and Southern New York. Sardeson (1924) and Allen (1929, p. 240; 1932) reported altered volcanic ash beds in Minnesota, Iowa, and Wisconsin. All these Eastern occurrences are found in the Trenton and Black River limestones (Middle Ordovician).

The Eastern beds were classed as bentonites because they occur as thin, widespread layers and the material slakes when placed in water. Ross (1928, p. 151) reported euhedral crystals of biotite and crystals of orthoclase. Volcanic structure is rarely preserved and has been noticed by Ross in only a few thin sections.

Since much of the clay has been metamorphosed by slippage along bedding planes, Ross (1928, p. 164) suggested it be called metabentonite (highly altered bentonite). However, as much of the material is not metamorphosed, it is referred to as potassium bentonite (K-bentonite) to differentiate it from the Ca, Na, Mg montmorillonitic bentonites. Bradley (1945, p. 711) proposed that the K-bentonite consists of a mixed layer mineral composed of illite and montmorillonite.)

Samples studied.—Twenty-two samples of K-bentonite were collected from 5 localities and 14 stratigraphically different beds in Central Pennsylvania.

The beds range from half an inch to 10 inches in width, but the majority are 2–4 inches. The color varies from tan to gray. Nearly all the beds contain abundant thin layers of secondary calcite.

Fresh K-bentonite is hard, has a semiconchoidal fracture, and is quite blocky. Solid pieces 4–6 inches thick and a foot square can be obtained. More weathered material has a shaly texture, breaks into thin chips, and grades into a highly weathered plastic clay with no evidence of shaly structure.

Figure 1, modified from Kay (1944, p. 3), is a columnar section of Middle Ordovician rocks in Central Pennsylvania. Beds of K-bentonite are indicated by numbers 0 to 7, and the letters N_1, N_2, and A to F. Rosenkrans (1933) described these beds in detail. Table 1 gives the loca-

tion and beds from which the present samples were obtained.

Table 2 is a list of the samples with brief megascopic description.

analyses. The logarithm, to the base 2, of the diameter is used, and the log is multiplied by -1 to avoid negative numbers: phi = $-\log_2$ diam. (mm).

TABLE 1.—LOCALITIES AND BEDS FROM WHICH K-BENTONITE SAMPLES WERE COLLECTED

Location		Symbol for Beds
Oak Hall	Oak Hall Quarry (OH)	2 0 N_1 N_2
Bellefonte	Bell Mine Quarry (B)	A D
Bellefonte	Bell Mine (M)	N_2 A D
Bellefonte	Water Quarry (W)	A B C D E
Salona	Salona Quarry (S)	5 4 3 2 1 0 A F

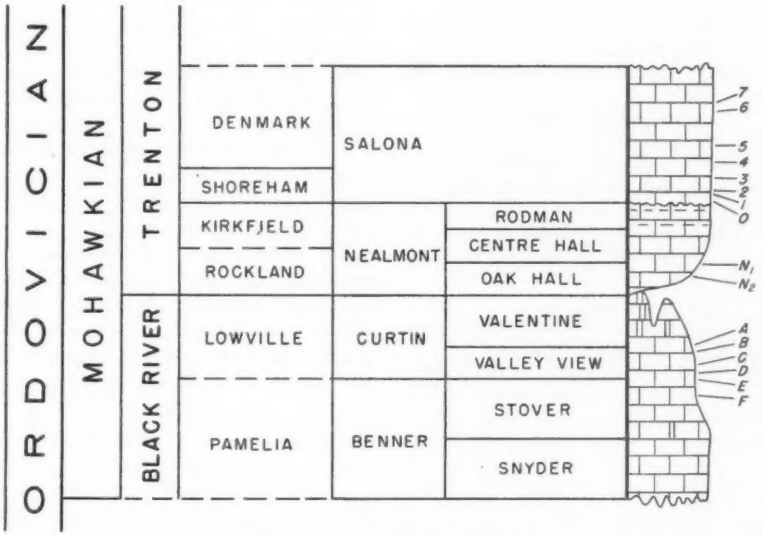

FIGURE 1.—PORTION OF THE MIDDLE ORDOVICIAN SECTION IN CENTRAL PENNSYLVANIA SHOWING THE LOCATION OF THE K-BENTONITE BEDS
After Kay (1944, p. 2).

MINERALOGY

Bulk Samples

Pipette analysis.—Twenty grams each of 16 of the K-bentonites were placed in 500 cc. of distilled water, 8 drops of NH_4OH were added, and the samples were shaken for 24 hours on a rotary shaker. The samples were then diluted to 1000 cc. and pipetted at 1 ϕ intervals. Cumulative curves were constructed (Fig. 2). The ϕ scale as derived by Krumbein (1934, p. 76) simplifies the calculations required in statistical

The cumulative curves give a graphic representation of the amounts of material in each grade size. As it is doubtful that the method used completely dispersed the clays, the curves are actually a measure of the degree of dispersion rather than of the size relationships. There are two types of curves. One is almost a straight diagonal line between 4 and 11 ϕ. The other shows a rather sharp increase in slope around 7 or 8 ϕ extending to 10 or 11 ϕ, which indicates that the unique frequency curve has a fairly sharp peak with a mode near 9–10 ϕ.

TABLE 2.—DESCRIPTION OF K-BENTONITE SAMPLES

Sample No.	Description
S-5	Light gray, highly weathered, plastic
S-4	Light gray, highly weathered, plastic
S-3	Light gray and light tan, moderately weathered, shaly
S-2	Tan, fresh, hard, blocky
S-1	Tan, highly weathered, plastic
S-0	Tan and gray, fresh, shaly
S-A	Gray, moderately weathered, shaly
S-F	Gray, moderately fresh, shaly
W-A	Light tannish gray, highly weathered, very fine chips
W-B	Light tannish gray, highly weathered very fine chips
W-C	Light tannish gray, moderately fresh, shaly
W-D	Gray, moderately weathered, shaly
W-E	Tan, moderately weathered, shaly
B-A	Gray, moderately weathered, fine chips
B-D	Gray, moderately weathered, shaly
M-N_2	Gray, fresh, very hard, blocky
M-A	Gray, moderately weathered, shaly
M-D	Gray, moderately weathered, blocky to shaly
OH-2	Tan, fresh, blocky
OH-O	Gray and tan, moderately fresh, shaly
OH-N_2	Gray, fresh, blocky to shaly
OH-N_1	Gray, fresh, blocky

which is the least weathered K-bentonite and yet produces a practically straight, horizontal curve. This is because only 32.5 per cent of the material was dispersed to finer than 4 ϕ (12.5 per cent between 4 and 11 ϕ) and the change in slope would not be apparent. M-N_2 has been slightly metamorphosed and as a result is much harder than any of the other K-bentonites.

When the percentage of material smaller than 1 micron is totaled for each sample the figures for the moderately weathered and highly weathered samples lie in a cluster of values ranging only from 19.5 to 20.3 per cent, whereas the moderately fresh and fresh K-bentonites have values ranging from 5.87 to 34.0 per cent.

The pH values of a dilute water solution of each sample were taken. The highly and moderately weathered samples have values ranging from 6.00 to 7.80; the moderately fresh and fresh specimens have values ranging from 4.4 to 7.95. Figure 3 illustrates, in the weathered samples, how the pH values and the amount of material in the less than one micron fraction tend to be clustered in a small area.

This would indicate that weathering tends to decrease the amount of variation in these properties of the K-bentonites, as is to be expected.

Differential thermal analysis data.—Differential thermal patterns were run on 22 of the bulk samples. The curves are similar to those

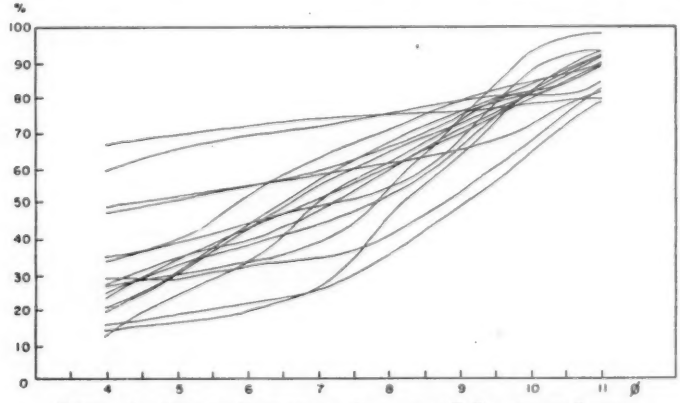

FIGURE 2.—SIXTEEN CUMULATIVE CURVES OF K-BENTONITE SAMPLES

The K-bentonites which have curves with a break in the slope are fresher than those with the constant slope except for sample M-N_2, produced by illite and montmorillonite. In these 2:1 minerals the first endothermic peak is caused by the loss of adsorbed water; the

second, in the 500°C to 800°C range, accompanies the loss of most of the water from the lattice, and the third endothermic peak is interpreted as being associated with the final destruction of the 2:1 structure. The high-temperature exothermal peak reflects the crystallization of spinel.

Two types of differential thermal curves (Fig. 4) were obtained. All curves show broad endothermic peaks at 100°–200°C. In many of the curves this is a double peak due presumably to the loss of water of hydration from the divalent adsorbed ion calcium. This peak is followed by a broad endothermic swell which generally starts near 500°C, slopes gradually to a maximum at 700°C, and returns to the base line rather sharply. However, eight of the curves (S-2, S-3, S-4, S-5, M-A, M-N_2, OH-2, W-C) show two peaks in this region, (1) at 600°C, (2) at 700°C. Samples S-4 and OH-2 have a stronger peak at 600°C than at 700°C.

All the curves have another endothermic peak between 900°C and 1000°C. Calcite impurities enlarge this peak or produce a second in this range.

Various exothermic peaks between 300°C and 600°C are probably due either to organic matter or iron sulphide or both. A distinct exothermic peak occurs in all the patterns at 1000°C or slightly higher.

Grim and Rowland (1942, p. 758), and Kerr, Kulp, and Hamilton (1949, p. 40) have reported curves similar to these in samples of K-bentonite from Kentucky and Virginia. Grim and Rowland call curves of this type a mixture of illite and montmorillonite and show that synthetic mixtures of these two minerals give such curves.

In the K-bentonite patterns the broad 100°–200°C peak is characteristic of both illite and montmorillonite, although the peaks appear to be too shallow for pure montmorillonite. The gentle swells and pronounced peaks near 600°C are generally considered characteristic of illite, and those at 700°C of montmorillonite. The 900°–950°C endothermic reaction is characteristic of both illite and montmorillonite, although the endothermic peak and the following exothermic peak at 1000°–1020°C are both quite high for illite and most of the montmorillonites. Although in general the curves are similar, there is considerable variation particularly in the critical 500° to 800°C range, and it seems likely that the K-bentonites are not uniform in chemical and/or mineral composition.

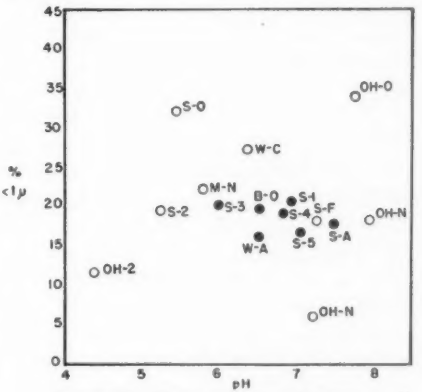

FIGURE 3.—PLOT OF pH VS AMOUNT OF DISPERSION

The figure illustrates how, in the weathered samples (filled circles), the pH values and the amount of material in the less than one micron fraction tend to be clustered in a small area.

X-ray data.—X-ray spectrometer and powder photographs were made of 21 of the bulk samples. Nearly all the patterns contained, in addition to reflections from clay minerals, a few lines due to calcite or quartz. The (001) line of the clay mineral found in each sample is given in Table 3.

Whereas in each complete pattern the majority of the *d* values check closely with those listed for illite there are many anomalous lines. One sample has a line with a value of 12.4Å. All the others contain a moderately strong (001) line with a value between 10.1 and 10.5Å. Eight samples give a very weak (001) line near 14Å. These are the same ones which have double hydroxyl-endothermal reactions (at 600° and 700°C).

Summary.—Examination of the bulk samples shows that weathering tends to restrict the amount of variation of the pH values and the dispersion characteristics of the K-bentonites. The differential thermal and x-ray patterns indicate a 2:1 clay with parameters similar to some of those of both illite and montmorillonite.

FIGURE 4.—DIFFERENTIAL THERMAL CURVES

1—Less than one micron fraction of K-bentonite sample OH-2. 2—Less than one micron fraction of K-bentonite sample OH-N_1. 3—Less than one micron fraction of sample OH-1B. 4—Insoluble residue from limestone sample OH-2A. 5—Insoluble residue from limestone sample OH-4C. 6—Bulk sample of Devonian "Tioga Bentonite".

Fine Fraction

Introduction.—On the basis of the bulk analyses the 22 clays were divided into two groups which differ in their x-ray and differential thermal characteristics. A typical sample (OH-2, OH-N_1) from each group was investigated in detail. The differential thermal curve of sample OH-2 has both a 600°C and 700°C hydroxyl-endothermal peak, whereas sample OH-N_1 has only the 700°C peak. X-ray patterns of OH-2 have 001 lines at 14.0Å and 10.6Å; OH-N_1 has only one line which is at 10.3Å. Both are fresh samples.

X-ray data.—Approximately 50 x-ray powder photographs were measured and studied in

TABLE 3.—X-RAY DATA ON BULK SAMPLES
001 Lines

Sample	d in Å	I	d in Å	I
S-O			10.2	m
1			10.3	m
2	13.9	ww	10.3	m
3	14.0	ww	10.2	m
4	13.9	ww	10.5	m
5	14.0	ww	12.4	w
A			10.3	m
F			10.2	m
M-A	14.0	ww	10.2	m
D			10.1	w
N_2	14.2	ww	10.3	m
B-D			10.2	m
OH-O			10.4	m
2	14.0	w	10.6	m
N_1			10.3	m
N_2			10.5	m
W-A				
B			10.4	m
C	14.0	ww	10.3	m
D				a
E			10.3	m

m = moderate; w = weak; ww = very weak; a = absent

clays was treated, and the measured values of the resulting 001 lines with the largest spacing.

The 060 value of 1.50Å (not given in Table 4) is characteristic of all patterns and indicates that the clay mineral is dioctahedral (contains 4Al in the octahedral position). The 001 values ranging from 10.3 to 10.9Å are not characteristic of any pure clay mineral but are commonly found for randomly interstratified 2:1 clay minerals.

Hendricks and Teller (1942, p. 150) devised a formula for predicting the 001 values to be expected from a random mixture of two sheet-like layers of different thickness. This formula was modified by Brown and MacEwan (1950, p. 241) who constructed a series of graphs from which the 001 spacings resulting from mixtures of two different sheet structures could be read.

Several factors determine the 001 values of randomly mixed layers of different thickness. The 001 values resulting from the mixture fall between the closely associated (001) values of the individual layers. The position of these peaks is modified further by the structure factor of the layers, by the polarization of the diffracted radiation, by the form of the specimen, and by the relative amounts of the two structures.

TABLE 4.—LARGEST 001 SPACINGS

	OH-2	OH-N_1	Montmorillonite
No treatment	10.9Å m 14.0Å ww	10.8Å m	
H	10.9 m 14.0 ww	10.8 m	14.5Å *
Ca	10.8 s 14.0 ww	10.9 m	15.1 *
Na	10.4 s	10.4 m	11.9 *
K	10.3 s 14.0 ww?	10.4 m	12.0 *
K (boiled 60 hours)	10.4 ss 14.2 ww?	10.3 s	
NH_4	10.3 s		12.1 *
E. glycol	9.5 m, 11.1 m	9.5 m, 11.3 m	17.0
24 hrs at 500°C	10.0 s 14.2 m		
24 hrs at 600°C	10.0 s 13.4 w		
24 hrs at 640°C	10.1 s 13.1 ww		
24 hrs at 675°C	10.0 s —		
24 hrs at 700°C	10.0 m —	10.1 s	

* Barshad (1950, 227)
ss = very strong; m = moderate; s = strong; w = weak; ww = very weak

detail. The interpretation of these patterns is based on the work of Bradley (1945; 1950) and Brown and MacEwan (1950).

Table 4 lists the various cations with which the less than one micron fraction of each of the

Table 4 shows that the untreated clay, the H based clay, and the Ca based clay all have

similar values of 10.8–10.9Å. Barshad (1950, p. 227) has reported that the 001 spacing for H montmorillonite is 14.5Å and for Ca montmorillonite 15.1Å. Assuming, therefore, a random mixture of nonexpanded 10Å dioctahedral 2:1 layers and either 14Å or 15.4Å expanded layers, the relative amounts of the two types of structures can be read from the graphs of Brown and MacEwan.

TABLE 5.—RANDOM LAYER DATA

Data from Brown and MacEwan (1950, p. 244–249)			K-bentonite
10/14 f 0.25	Original Indices	10/15.4 f 0.20	Untreated H, Ca
10.9s	001/001	10.9m	10.8–10.9m
4.90m	002/003	5.08s	4.94–5.00m
3.37m	003/004	3.31m	3.25 to 3.43m broad
10/12.4 f 0.25		10/12.4 f 0.20	Na, K, NH_4
10.5s	001/001	10.3s	10.3–10.4s
5.14w	002/002	5.12w	5.11–5.19w
3.26m	003/004	3.31m	3.27–3.30s
10/17.7 f 0.25		10/17.7 f 0.20	Ethylene glycol
11.7*	—/001	11.4*	11.1–11.7*
9.5m	001/002	9.6m	9.5–9.7m
4.95 to 5.28 broad	002/003 w	5.03w	4.67 to 5.13w broad
3.42w	003/005	3.37w	3.34s

— = range; * = diffuse; f = proportion of layers with higher spacing; 001/001 = nonexpanded indices/expanded indices; m = moderate; w = weak; s = strong

Table 5 shows the three values taken from the graphs and the 001 lines of the individual structures which combine to give the resulting 001/001 value of the mixed layer mineral; (f = the proportion of layers with higher spacing). For a mixture of 10Å and 14Å layers an f of 0.25 gives a value equal to the measured value. For a mixture of 10Å and 15.4Å layers an f of 0.20 produces the same result. Brown and MacEwan list three 001/001 values for each mixture. Table 5 shows that all three figures for each f value check closely with the figures obtained by measurement of the K-bentonite patterns.

These data indicate that the clay contains 75–80 per cent of 10Å layers and 20–25 per cent of expanded layers. The expanded layers have a thickness of between 14 and 15.4Å.

Table 4 shows that the Na, K, and NH_4 based K-bentonite sample give 001/001 values of 10.3-10.4Å. Barshad (1950, p. 227) lists the 001 values for montmorillonite treated with these cations as: Na = 11.9; K = 12.0; NH_4 = 12.1. On the basis of the above work, using an f value of 0.20 to 0.25 and end members of 10Å and 12.4Å, the calculated results should check with the measured values if the previous interpretation is correct. As shown in Table 5, the 001/001 value for 10/12.4 with f = 0.20 is 10.3 and for f = 0.25 is 10.5; the measured values are 10.3–10.4. The other two 001/001 spacings also check with the calculated values.

A further check can be made by soaking the clay in ethylene glycol and expanding the expandable layers to 17.0Å. The results are shown in Table 5; the measured and calculated values are quite similar. The 11.4–11.7Å calculated values are broad and extend upward to larger values. Also, whereas in the other graphs the principal 001/001 values were obtained by a combination of the 001 spacings of both structures, in the 10/17.7 spacings the resulting value is formed by a combination of the 001 reflection from the 10Å unit, and the 002 from the 17.7Å unit (8.85Å) giving 001/002 value less than 10Å. This is probably the origin of the 9.5Å spacing which Jackson, Hseung, Corey, Evens, and Vander Heuvel (1952, p. 3) attribute to 003 of a 28.5Å (10 + 18.5Å) spacing.

A broad spacing extending from 4.95 is calculated for the f = 0.25 value. In some of the measured patterns this line split into two distinct lines at 4.97Å and 5.25Å. The calculated values also indicated two distinct lines as f is increased.

These x-ray data show that sample OH-N_1 is composed of randomly interstratified dioctahedral layers of which 75–80 per cent are nonexpanded and 25–20 per cent are expanded. Sample OH-2 has the same types of layers mixed in the same relative proportions plus some packets of a mineral with a 14Å 001 spacing.

Table 6 contains a list of all the mixed layer lines measured from a typical x-ray pattern of OH-2, together with the lines for illite from Calhoun County, Illinois, as given by Grim, Bray and Bradley (1937, p. 821). The table also contains the indices for muscovite and the randomly interstratified mineral. All reflections of the mixed layer material except the 001 series are similar to those for illite.

TABLE 6.—MIXED LAYER SPACINGS AND INDICES

Samples			Indices		
OH-2 H clay d in A	I	Illite Calhoun Co. d in A	I	Muscovite	Randomly interstratified mineral
10.9	m				001/001
		9.98	s	002	
		4.97	w	004	
4.95	m				002/003
4.47	m	4.47	s	110	
		4.11	ww	002	
		3.7	ww	023	
3.43	m	3.4	ww	114	
3.35	m	3.31	m	006	003/004
3.25	m	3.2	ww	114	
3.11	ww			$11\bar{5}$	
		2.98	w	025	
2.83	w	2.84	ww	115	
2.55	m	2.56	s	$20\bar{2}$	
2.495	w			008	004/005
2.45	w	2.44	w	$13\bar{3}$	
2.38	mw	2.38	m	133	
2.245	w	2.24	m	$22\bar{1}$	
2.18	w	2.18	w	$22\bar{3}$	
2.13	w	2.11	w	043	
1.97	m	1.98	m	0010	005/007
1.65	mw	1.65	w	310	
		1.64	m	312	
1.50	m	1.50	s	060	
1.42	ww			0014	
1.34	w	1.34	ww	335	
1.29	w	1.29	m	400	
1.265	w			402	
1.245	w	1.24	w	0016	

s = strong; m = medium; w = weak; ww = very weak

Table 4 shows values of 14.0Å in sample OH-2 which are not present in sample OH-N_1. This spacing is due to packets of 14.0Å layers interspersed among, but not as part of, the predominant interstratified layers.

Sample OH-1B, found immediately above Sample OH-2, gave an x-ray pattern containing a 14Å line three to four times as dark as that in any of the previously examined sample.

TABLE 7.—DATA ON 001 VALUES FOR HEATED CHLORITE

	Brindley and Ali	OH-2	OH-1B
temperature	001	001	001
20°C to 300°C	14.45Å	14.0Å	14.0Å
600°C	14.22	13.4	13.6
660°C	14.09	13.1	

Using the data of Brindley and Ali (1950, p. 26) it was possible to prove that the 14Å reflection was caused by a chlorite mineral. These authors reported that, when powdered penninite is heated to 600°C for 2 hours, the x-ray intensities change. They calculated x-ray intensities for various atomic arrangements, using a one-dimensional Fourier analysis of the electron density distribution. The best agreement between calculated and measured intensities occurred when they assumed two-thirds of the brucite layer to be decomposed and the liberated Mg and O atoms to be in the hydroxyl sheets. As the chlorite is heated the first stage of dehydration consists of a loss of about half of the eight water molecules at 600°C. This is accompanied by a migration of the liberated Mg atoms into or toward the hydroxyl sheets of the brucite layer, and the original 14.45Å lattice height is reduced to 14.22Å. At 660°C the remaining water is lost, and the lattice shrinks to 14.09Å. At 760°C the structure breaks down, and most of the penninite lines disappear.

Table 7 shows the 001 values obtained by Brindley and Ali (1950, p. 27) for chlorite heated for 2 hours and the values obtained for samples OH-2 and OH-1B heated for 24 hours.

The differences between the chlorite packets of the K-bentonite samples and the chlorite of Brindley and Ali may be due to inaccurate measurement and to the length of the heat treatment, but more probably indicate that the chlorite in the K-bentonite is a different variety from the pure Mg chlorite of those authors.

Table 8 shows the relationship between the observed intensities at normal temperature and at 600° as described by Brindley and Ali (1950, p. 26) and those for samples OH-1B.

TABLE 8.—CHANGE OF X-RAY INTENSITIES OF THE 00l SERIES OF CHLORITE HEATED TO 600°C

00l	Brindley and Ali		OH-1B	
	Normal	600°C	Normal	600°C
001	24	121	m	s
002	67	23	wm	a
003	81	30	m	ww
004	100	100	wm x	ww
005	67	186	m x	w
006	0	100	w x	a
007	37	58	w	ww
008	11	0	a	a
009	11	43	w x	ww
0010	37	62	ww x	a
0011	0	40	a	a
0012	33		a	a
0013	0		a	a
0014	27		a	a

s = strong; m = moderate; w = weak; ww = very weak; x = overlapped by lines of other minerals; a = absent

The comparison is not exact as the chlorite is probably not a pure Mg variety and as the x-ray patterns of OH-1B contain overlapping lines for a mixed-layer mineral and for quartz. However, the sharp increase in intensity of the 001 line and the definite weakening of the 002 and 003 lines when the material is heated to 600°C strongly suggest that the mineral is a chlorite.

In Table 9 are listed the *d* values for OH-1B and the range of *d* values for 20 chlorite patterns taken from the literature. Many of the lines are overlapped by lines of other minerals, but the unambiguous lines check quite closely with the chlorite values. Particularly important is the 1.55Å 060 reflection which is characteristic of a trioctahedral mineral.

Differential thermal analysis data.—Differential thermal analyses were made of the less than one micron fractions of the untreated, the electrodialized, and the KOH, NaOH, and $CaCl_2$ treated clays.

Differential thermal curves of untreated clays OH-2 and OH-N_1 are shown in Figure 4.

TABLE 9.—CHLORITE X-RAY DIFFRACTION DATA

Literature *d*	Intensity	OH-1B *d*	Intensity	Index	Interfering lines
13.1–15.0Å	w-m	14.0Å	m	001	
6.7–7.2	s	7.1	w-m	002	
4.5–4.8	w-s	4.7	m	003	
3.92	w	3.87	ww		M
3.42–3.58	m-s	3.52	wm	004	
3.13	ww	3.13	ww		M
2.83	m	2.83	m	005	M
2.70–2.79	ww-m	2.75	ww		qtz.
2.55–2.59	m	2.57	s		
2.38–2.46	m	2.45	w		qtz. M
2.33	ww	2.34	w	006	M
2.25–2.27	w	2.24	m		M
2.00–2.09	w-m	2.03	w	007	
1.96–2.00	ww-m	2.03–1.97	w		M
1.88–1.90	ww-m	1.86	w		
1.73–1.76	ww				
1.66–1.68	w	1.67	w		
1.55–1.58	w-s	1.55	w	009, 060	
1.51–1.53	w-s	1.51	w		
1.41–1.44	ww	1.42	ww	0010	M
1.37–1.39	w				
1.30–1.33	w	1.29	m		M

M = mixed layer clay lines; qtz. = quartz lines; s = strong; m = medium; w = weak; ww = very weak

All patterns contain a strong endothermal peak at 100°–150°C. In addition the curves of the $CaCl_2$ treated clays contain a small endothermal peak at 225°C which was found elsewhere only in curves of the untreated clays. This would suggest the presence in the original clay of some Ca which was replaced, in the other samples, by the K, Na, and H cations.

The hydroxyl-endothermal peaks were not shifted in position although in OH-2 the amplitude of the 700°C peak relative to the 600°C peak tended to increase in the samples treated with K, Na, and Ca.

OH-2 was electrodialized for 300 hours removing some of the iron and magnesium and possibly some aluminum from the structure. This lowered the amplitude of the 600°C peak

considerably but did not affect the 700°C peak, thus indicating preferential removal of iron and magnesium from the layers responsible for the lower temperature peak.

Examination of the differential thermal patterns of the 22 samples showed that 8 of the patterns contained a 600°C peak, usually very small. The x-ray diffraction patterns were re-examined, and several new x-ray pictures were taken using a longer exposure. All samples which showed a 600°C differential thermal peak gave a weak chlorite x-ray pattern. It was necessary to determine if the layers which were responsible for the 600°C hydroxyl-endothermal reaction were 14Å chlorite layers.

Sample OH-1B, which gave an x-ray pattern containing a 14Å chlorite line three to four times as dark as that in any of the previous samples, has a differential thermal curve (Fig. 4) with a sharp 600°C endothermal peak approximately three times as large as that in sample OH-2. The endothermal reaction is accompanied by a lowering of the lattice height to 13.8Å indicating that this peak is at least partly associated with a loss of water from the chlorite layers.

A review of the literature reveals that most chlorites have a strong endothermal peak at 550–600°C and a smaller one at 800°C, followed by a weak to moderate exothermal peak. OH-1B, in addition to the strong endothermal peak at 600°C, has a weak one at 825°C followed by a weak exothermal peak (Fig. 4). These peaks are strikingly similar to that shown by Barshad (1948, p. 667) for prochlorite from Chester, Vermont. The data indicate that the chlorite packets in the K-bentonite samples are responsible for the 600°C endothermal peaks, and the randomly interstratified units, low in iron, give the single peak at 700–725°C. The literature also reveals that most illites contain 5–10 per cent iron oxides. The presence of considerable iron in montmorillonite lowers the endothermal peak to about 500°C. These K-bentonites contain very little iron (0.31 and 1.28 per cent), and this may be responsible for the high temperature of the hydroxyl-endothermal reaction.

It has commonly been stated that the double hydroxyl-endothermal peak found in patterns of mixed-layer clays is caused by the two individual clays forming the mixture, the illite layers causing the lower peak, the montmorillonite layers the higher. However, the present study indicates that it is not necessary to have two hydroxyl-endothermal reactions occurring in a mixed-layer clay. It is believed that the single hydroxyl-endothermal peak in the curve of sample OH-N_1 is characteristic of the mixed-layer mineral in the K-bentonites.

A possible explanation can be obtained from the chemical-analysis data which suggest that the 80 per cent nonexpanded layers may contain as much as 9.2 per cent K and would therefore be similar to muscovite. Fine-grained muscovite gives a broad, shallow hydroxyl-endothermal reaction extending from 475° to 775°C (Barshad, 1950, p. 233). If these nonexpanded layers are sufficiently muscovite-like they may cause the gradual endothermal reaction which starts at approximately 500°C and extends to higher temperatures. The hydroxyl-endothermal peak at 700°C would be caused by the 20 per cent expanded, montmorillonite-like layers.

Chemical data.—The chemical analyses listed in Table 10 were made in an attempt to find the relation of the structural formulas of these clays to others in the literature and to see what differences between the two clays were caused by the presence of chlorite in OH-2. X-ray diffraction patterns of sample OH-2 show a very weak 14Å line and a strong 10.9Å line. OH-N_1 contains only the strong 10.9Å line.

The low percentage of chlorite has not affected the chemical analyses appreciably. However, the slightly greater amounts of FeO and MgO and the lesser amount of SiO_2 in the OH-2 sample as compared with the OH-N_1 sample can probably be explained by the presence of a small percentage of a chlorite mineral.

The most noticeable difference between the K-bentonite and the illite analyses is the much smaller amount of iron in the former. This is believed to be the reason for the higher temperature (700°C) endothermal peak in the K-bentonites. It is of interest to note that OH-N_1, which has less iron than OH-2, has its endothermal peak occurring approximately 25° higher.

The following structural formulas were com-

TABLE 10.—CHEMICAL ANALYSES

	OH-2*	OH-N_1*	Illite†
SiO_2	50.65	52.44	48.95
Al_2O_3	27.37	26.38	25.03
Fe_2O_3	1.15	.31	7.29
FeO	.13	.00	1.61
MgO	3.75	3.57	3.10
CaO	.73	.66	.29
Na_2O	.16	.16	.15
K_2O	5.89	7.39	6.03
H_2O+	5.73	4.78	7.56
H_2O-	3.47	3.07	1.70
CO_2	.06	.00	
TiO_2	.35	.44	.51
P_2O_5	.03	.04	
SO_3	.01	.30	
Cl	.01	.02	
MnO	.002	.001	
Li_2O	.01	.01	
Total.....	99.50	99.57	102.22

* OH-2 and OH-N_1 less than one micron fraction: Analyzed by J. A. Maxwell, Department of Geology and Mineralogy, Univ. of Minn.

† Illite: Average of five analyses (Grim *et al.*, 1937, p. 823)

puted from the chemical analyses using the method of Ross and Hendricks (1945, p. 42).

$$K_{.55}\frac{Ca}{2}\ .11\ Na_{.02}$$

OH-2: $(Al_{1.68}Fe'''_{.06}Fe''_{.01}Mg_{.25})(Al_{.55}Si_{3.45})O_{10}(OH)_2$

$$K_{.64}\frac{Ca}{2}\ .10\ Na_{.02}\ \frac{Mg}{2}\ .04$$

OH-N_1: $(Al_{1.64}Fe'''_{.02}Mg_{.34})(Al_{.45}Si_{3.55})O_{10}(OH)_2$

The presence of the chlorite is indicated by the total of 2.10 in octahedral co-ordination for OH-2 as compared to 2.02 in OH-N_1. If the formula for OH-2 is corrected for the presence of a small amount of chlorite by subtraction of small quantities of magnesium and iron and the addition of some silica, the two formulas will be in close accord.

In OH-N_1, if the .02 excess magnesium in the octahedral layer is assigned to an exchange position (Foster, 1950, p. 1462), the excess positive charge on the octahedral and tetrahedral layers would total .79, and the total negative charge of the external cations would be .80.

Most, if not all, of the K is probably present in the nonexpanded layers, whereas the Ca, Na, and Mg are likely held on the expanded layers. A direct relationship probably exists between the amounts of K and Ca, Na, and Mg present and the degree of expandability, as these latter three elements make up 20 per cent of the total external cations and the x-ray reveals that the expanded layers form 20–25 per cent of the total clay. As 80 per cent of the external cations are K and as 75–80 per cent of the layers are nonexpanded, undoubtedly the K prevents expansion. This close correlation between ratios obtained by x-ray and chemical data suggests that the proportion of mixed layers can be predicted from the chemical data, or in turn the composition of the external cations can be predicted from the x-ray data.

If all the K is contained in the 80 per cent nonexpanded layers, these layers would contain 9.2 per cent K—enough to have them classed as muscovite. It is impossible to determine if the expanded and nonexpanded lattices have different compositions.

On the basis of the x-ray and chemical data the K-bentonite can be considered a clay composed of randomly interstratified expanded and nonexpanded dioctahedral 2:1 layers. These layers are in the proportion of 1:4 (KCa, Al tetoctrite)[1].

Electron microscope data.—Electron micrographs of all 22 clay samples were quite similar and Figure 1 of Plate 1 is of the less than one micron fraction of sample OH-N_1. The electron micrographs show that the clays consist of thin, irregular, subequant flakes which have smooth, straight edges.

Electron micrographs were taken of the clay samples after each of the chemical treatments and none showed any significant variation in morphology.

Summary.—The investigation of the less than one micron fraction of these two samples shows that the K-bentonite beds are composed largely of an expanded and nonexpanded dioctahedral 2:1 clay mineral, in the ratio of 1:4 and that many of these beds contain admixed packets of chlorite. The chlorite is responsible for the

[1] Terminology after Weaver (1953).

600°C hydroxyl-endothermal peak, and the mixed layer mineral the 700°C peak. The nonexpanded layers contain K as their external cation, and the expanded layers contain Ca, Na and Mg. The nonexpanded layers may contain as much as 9.2 per cent K and might therefore be classed as muscovite.

Petrography

X-Ray Analysis

X-ray spectrometer patterns were made of the bulk samples of all 22 clays. Nine thin sections and the heavy- and light-mineral separates of samples OH-2 and OH-N_1 were examined.

The heavy minerals were concentrated by crushing 30 gms. of clay, dispersing it in water, and then shaking it for 24 hours on a rotary shaker. The less than 15 micron fraction was decanted five times; the residue was dried and sieved through a 100-mesh sieve; the material passing through the sieve was placed in bromoform (Sp. Gr. 2.88) and centrifuged at 1000 RPM for 15 minutes; the lower portion of the bromoform, containing the heavy minerals, was frozen, and the remaining bromoform and light minerals poured off. The frozen bromoform was then thawed, and the heavy minerals collected in a filter paper. A clean separation was obtained by this method.

The bulk x-ray spectrometer patterns gave, in addition to peaks for clay minerals, very small quartz peaks indicating less than 5 per cent of this mineral.

Sample OH-2

In thin section, the clay constitutes approximately 95 per cent of the material present. The oriented aggregates show a white to yellow birefringence and have an index of refraction between 1.54 and 1.55.

The varietal materials, constituting approximately 5 per cent of the sample, are albite quartz, volcanic glass, apatite, zircon, biotite, and leucoxene.

Albite (Pl. 1, fig. 2): A few albite peaks show up on x-ray spectrometer patterns of the light fraction. In loose grain mounts, angular, xenomorphic grains are seen. The extinction angle on albite twinning, when present, ranges from 12° to 17°. The grains average .05 to .1 mm. in size and range up to 0.2 mm. They commonly contain minute calcite inclusions. Most of the grains have a rough, deeply pitted surface. X-ray patterns of the feldspar are similar to those for low-temperature albite listed by Tuttle and Bowen (1950).

Table 11.—X-Ray Data on Apatite from Sample OH-2

d	*I*
8.20Å	ww
4.51	ww
4.10	ww
3.45	m
3.34	ms
3.11	ww
2.82	s
2.73	ms
2.64	ww
2.54	ww
2.445	ww
2.395	ww
2.350	ww
2.270	w
2.151	ww
2.085	ww
1.949	ww
1.903	w
1.848	ww
1.809	ww
1.781	ww
1.758	ww
1.731	ww
1.724	ww
1.677	ww

s = strong; m = medium; w = weak; ww = very weak

Quartz: The small amount of quartz indicated by the x-ray patterns occurs as disseminated, fine-grained chert.

Volcanic glass (Pl. 1, fig. 3): In the thin sections and in the loose mineral mounts, about 3 per cent of the material consists of clear, isotropic, angular grains with an index of refraction of 1.55. Many grains are elongated and have a shard-like appearance.

Apatite (Pl. 2, fig. 1): Apatite grains constitute 90 per cent of the heavy-mineral separate and about 0.1 per cent of the total sample. The apatite occurs in two forms. The most common type, which constitutes 90 per cent of the grains, occurs as prismatic,

idiomorphic to hypidiomorphic grains ranging from .01 to .16 mm. in length. They average .08 mm. in length and .02 in width. The prism sides are smooth and straight; the ends are sometimes fractured, but usually contain basal or pyramidal terminations (many appear slightly rounded). Cross sections show a perfect hexagonal outline. Elongated bubble inclusions are common and usually occur in the center of the grain. The less common type of apatite consists of xenomorphic grains equant in shape, with rough, irregular edges. The surface of the grains is pitted, rough, and grayish. A few contain irregular cracks. Table 11 lists the *d* values of the apatite concentrate from OH-2. The *d* values and intensities do not check exactly with any of the apatites in the literature.

ZIRCON (Pl. 2, fig. 1): Zircon makes up 2 per cent of the heavy-mineral separate. Most of the zircon is prismatic and idiomorphic. A few crystals are fractured. The grains are colorless, and many contain large bubble inclusions. The average length is about 0.04 mm.

BIOTITE: A few flakes and books of light pinkish-brown biotite can be seen. The flakes vary from hypidiomorphic hexagonal to xenomorphic. Rod-shaped inclusions are common and generally lie at an angle of 60° to one another.

LEUCOXENE: About 10 per cent of the heavy-mineral separate consists of opaque, irregular, equant grains which are dull white in reflected light. X-ray patterns indicate the leucoxene is composed of anatase.

Sample OH-N$_1$

In thin section the clay is similar to OH-2. However, there are a few lenses of clay, within the main mass, which are characterized by disseminated, opaque, black material—probably pyrite.

No feldspar or quartz x-ray lines were obobtained, and no trace of these minerals or volcanic glass was seen under the microscope. The heavy minerals—apatite, zircon, and biotite—are identical in kind and relative amount, with those in OH-2. In addition, the heavy-mineral separate contains about 50 per cent of opaque minerals consisting of aggregates of pyrite and chalcopyrite.

ORIGIN

In considering the origin of the Ordovician K-bentonites, the data presented may be evaluated with regard to the probable nature of the source material, the manner of its deposition, and the mode of formation of the clay minerals.

The presence in the K-bentonite beds of volcanic glass, biotite, and idiomorphic apatite and zircon crystals supports the theory that the source material was volcanic. The wide lateral extent of the beds and the nature of the associated rocks indicate that the volcanic detritus was probably fine glass fragments laid down in shallow seas which were depositing large amounts of limestone. The lack of quartz phenocrysts and the fairly high index of refraction (1.55) of unaltered glass fragments (characteristic of a glass low in silica) plus the presence of the minerals enumerated, indicate that the ash probably had the composition of a trachyte or possibly a latite. However, it is likely that the unaltered glass has a composition different from the great mass of altered glass.

Presumably this material was the source of the mixed-layer clay mineral and chlorite. The presence of chert in the outer edges of the beds and in the adjacent limestone indicates that the volcanic detritus was also the source of silica.

Assuming that the relatively stable Al_2O_3 was not taken into solution, so much of the silica, iron, and alkalis of the volcanic ash was removed that about 1½ parts of glass were required to produce 1 part of clay. Some of the silica was precipitated to form the chert commonly found in the upper and lower portion of the bentonite beds.

After removal of sufficient amounts of the various ions, an impure pyrophyllite lattice was probably formed with a composition similar to that of the Cretaceous and Tertiary expanded

PLATE 1.—SOME MINERALS FOUND IN K-BENTONITES

FIGURE 1. ELECTRON MICROGRAPH OF SAMPLE OH-N$_1$
Well-developed, thin flakes with sharp edges. The less than one micron fraction.

FIGURE 2. ALBITE FROM SAMPLE OH-2
Large, pitted grains.

FIGURE 3. VOLCANIC GLASS IN SAMPLE OH-2
Note characteristic shard-like shape.

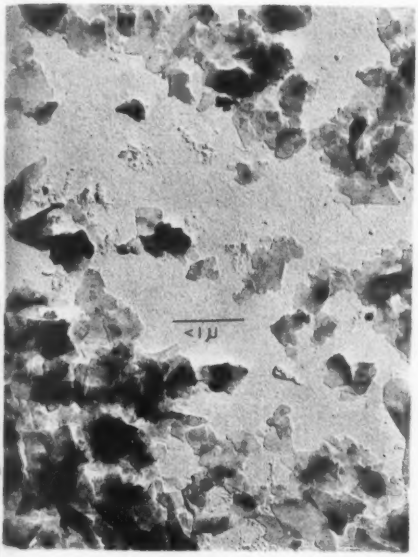

FIGURE 1

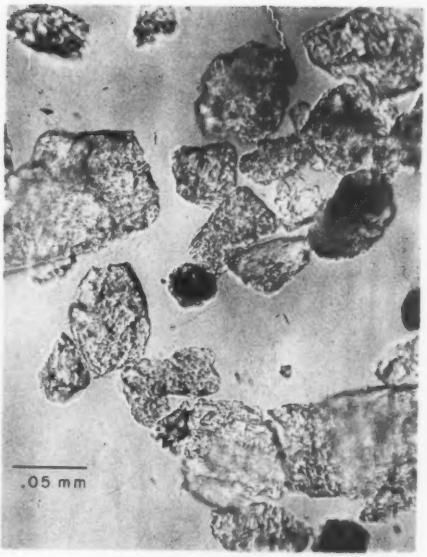

FIGURE 2

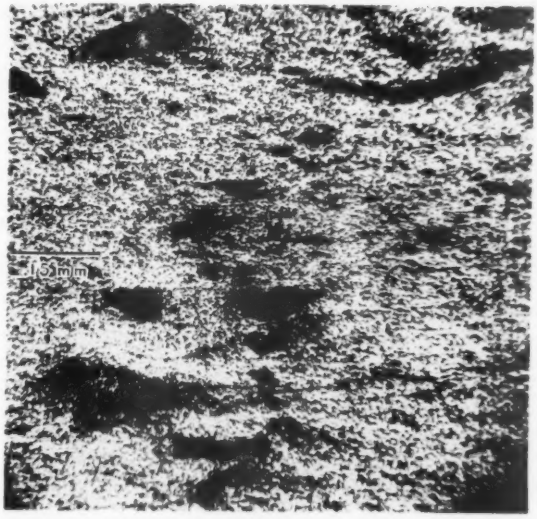

FIGURE 3

SOME MINERALS FOUND IN K-BENTONITES

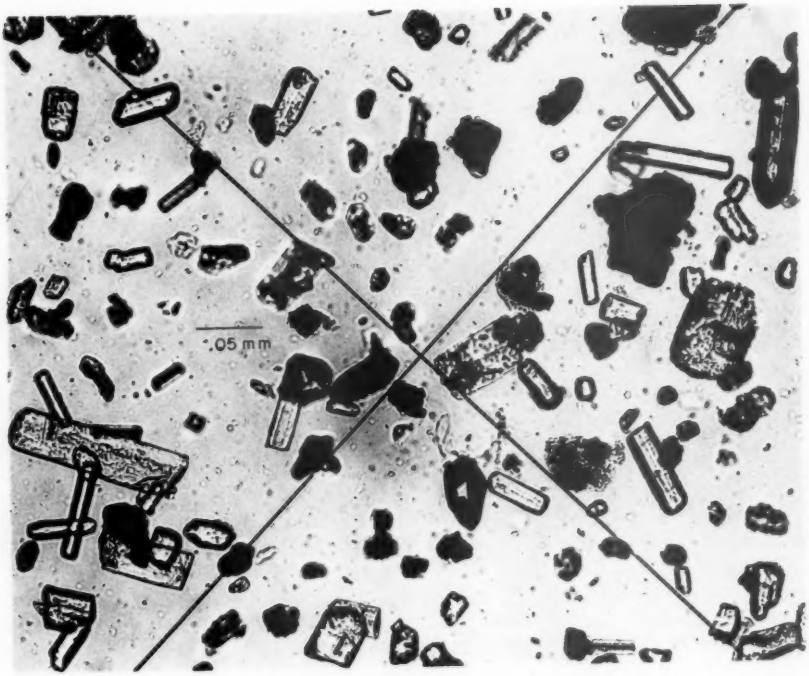

FIGURE 1

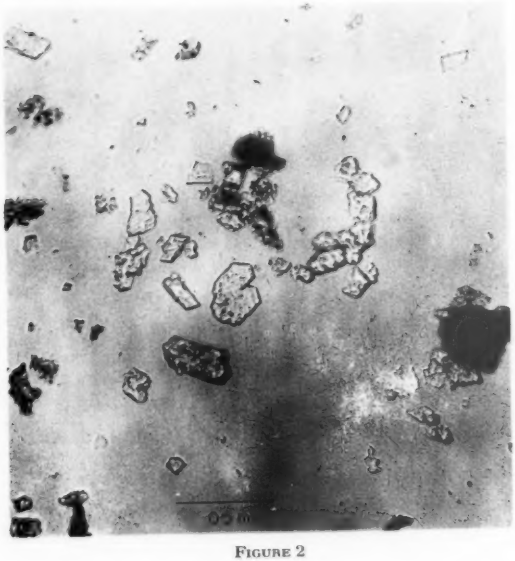

FIGURE 2

HEAVY MINERALS IN K-BENTONITE AND ALBITE IN LIMESTONE RESIDUE

dioctahedral 2:1 clays (bentonites). This is suggested by the high-temperature hydroxyl-endothermal reaction of the randomly interstratified clay mineral and is further verified by chemical analyses which, when presented in the form of structural formulas, are similar to many of those listed by Ross and Hendricks (1945, p. 42–43) for the montmorillonite-beidellite series. Many of the Cretaceous and Tertiary clays, which were derived from glass similar in composition to the Ordovician ash, were formed in saline lakes and seas. However, any potassium in the glass was removed, and the resulting clays attracted calcium as the most abundant external base. A similar situation probably prevailed in the calcium-saturated Ordovician sea suggesting that the K-bentonite, in its original form, probably had this ion as the predominant external base and consequently consisted of an expanded dioctahedral 2:1 clay. As the Ordovician period was relatively quiescent as compared to the Cretaceous and Tertiary, the Ordovician volcanic ash was probably in contact with sea water much longer than were the more recent ash falls. Therefore, it is believed that due to the availability of potassium in the sea water, accumulative retention occurred and extended over such a period of time that the calcium in the Ordovician K-bentonite was replaced in 80 per cent of the layers resulting in this proportion of a nonexpanded clay mineral.

Ordovician Limestone

Introduction

Samples of limestone above and below several of the K-bentonite layers were collected at Oak Hall Quarry 3 miles east of State College. The clay minerals and the varietal minerals in the limestone were compared to those in the K-bentonites to attempt to determine the duration of the Ordovician volcanic activity.

From the north wall of the quarry 26 samples were collected within a vertical range of 58 feet from three vertical sections (A, B, C) 25 feet apart. Within this range occur three beds of K-bentonite from which samples OH-2, OH-1, and OH-0 were obtained (Table 12). The limestone is a hard, dense, fine-grained, light-gray to black rock which in most cases is massive. It has a semiconchoidal fracture but immediately above and below the K-bentonite beds has a shaly appearance.

Carbonate Minerals

X-ray spectrometer patterns of six selected bulk samples (2A, 6A, 10A, 8B, 9B, 4C) of the limestone contained strong calcite peaks and no dolomite peaks, indicating a fairly uniform $CaCO_3$ limestone.

The calcite occurs as fine-grained xenomorphic crystals without noticeable orientation. The thin sections commonly show a nodular structure with the individual nodules 0.05–0.10 mm. in diameter. These subspherical granules consist of very fine-grained calcite particles, averaging 0.007 mm. in diameter, that are much darker than the matrix limestone which is quite clear. This dark color is apparently caused by clay concentrated in the nodules. Grains of the matrix calcite range from .015 to 0.50 mm. in diameter with the average tending toward the small size. Occasionally a few idiomorphic rhombs of dolomite are found.

Insoluble Residue

Description.—The limestone samples were crushed to pieces a fourth to half an inch in diameter and placed in a dilute solution of HCl until all the calcite was dissolved. The per cent of the insoluble residue for each sample is listed in Table 12. Laterally, within beds, there is little variation; vertically the percentage varies from 4.2 per cent 9 feet below K-bentonite bed OH-1 to 81.2 per cent immediately above bed OH-2. There appears to be an increase in the amount of insoluble residue in the

Plate 2.—Heavy minerals in K-bentonite and albite in limestone residue

Figure 1. Heavy minerals from Sample OH-2
Apatite and zircon.

Figure 2. Albite from Limestone Sample OH-4C
Idiomorphic, authigenic crystals form about 60 per cent of insoluble residue.

TABLE 12.—LOCATION OF OAK HALL SAMPLES

Feet	Sample	Section A % Residue	Section B % Residue		Section C % Residue		
0	0A	23.8					
	1A	9.8					
10	2A	23.8					
			0B	33.2			
	3A	40.8					
	4A	81.2					
	5A	100.0	1B	100.0			OH-2
20	6A	68.6	2B	70.8			
	7A	29.6	3B	32.0			OH-1
	8A	60.0	4B	52.0			
	9A	40.6					
			5B	9.6			
	10A	16.2	6B	7.2			
30							
			7B	27.0	0C	4.2	
			8B	10.6	1C	29.6	
							OH-0
			9B	29.0			
40					2C	22.2	
					3C	17.6	
					4C	10.4	
50							

limestone as the K-bentonite beds are approached. This would indicate that much of the residue was formed from volcanic detritus. The volcanic deposition began slowly, built up to a maximum when the K-bentonite beds were deposited, and then gradually fell off to a minimum before increasing again. In order to see if this was the actual situation the residues were examined to try to determinine their source.

Using the data from x-ray spectrometer patterns of the residues, seven samples were selected, and x-ray powder photographs taken. Further detailed studies were made on the less

than one micron fraction of three representative samples. Loose grains of each residue were examined under the binocular and petrographic microscopes, and in three samples (2A, 1B, 4C) the heavy minerals were separated and examined. Several thin sections were examined under the microscope.

All residues are dark black and contain considerable organic material. Table 13 lists the intensities of the peaks in the x-ray spectrometer patterns of the bulk, uncleaned residues. All samples show a prominent 10Å peak caused by a nonexpanded dioctahedral 2:1 clay (illite). In addition many of the samples give patterns with peaks at 14Å and 7Å which may indicate chlorite or an expanded dioctahedral 2:1 clay.

Light minerals.—The light minerals found in the insoluble residue are quartz, chert, and albite.

QUARTZ: All x-ray patterns show a strong quartz reflection. The quartz occurs in three forms:

(1) Detrital quartz. A few subrounded, fine (0.05 mm.) grains of detrital quartz are present.

(2) Authigenic quartz. Idiomorphic quartz crystals, ranging in length from 0.15 to 0.30 mm., are found in nearly all the limestone samples. The crystals have prism and rhombohedron faces. All edges have a rough, irregular contact with the calcite. Calcite inclusions are abundant, frequently making up half the quartz crystal.

(3) Chert. Chert commonly occurs in irregular fragments averaging about 0.05 mm. in size. The individual microcrystals are a few microns in diameter. The irregularly shaped angular grains suggest that the chert is authigenic.

A semiquantitative examination of the x-ray spectrometer patterns indicates a concentration of quartz near the bentonite beds.

ALBITE: Authigenic albite is found in every limestone residue. It is uniform in morphology throughout the three sections. Approximately 50 per cent of the grains are idiomorphic, and the majority of the others are hypidiomorphic. Twinning is not common. The majority of the grains are clear, contain no inclusions, and show no alteration. However, many of the grains have carbonate inclusions covering from 1 to 25 per cent of the surface of the crystal (Pl. 2, fig. 2).

The majority of the crystals fall in the 0.01–0.04 mm. range although some grains are as large as 0.08 mm. These larger grains are xenomorphic and angular with a rough, "pot-hole" surface and resemble the large crystals in the K-bentonite (OH-2).

Table 13 shows the relative strength of the albite peaks in the x-ray spectrometer patterns in each residue. The samples with the smaller

TABLE 13.—X-RAY SPECTROMETER DATA ON BULK LIMESTONE RESIDUES

	Quartz	Albite	10Å Illite	14Å Chlorite	Per cent residue
0A	m	m	m*		24
1A	m	m	m*		10
2A	m	w	m*		24
3A	s	w	w		41
4A	s	w	s	w	81
5A	s	ww	s	w	100
6A	s	ww			69
7A	ss	s	w		30
8A	s	m	ms		60
9A	s	w	ms	w	41
10A	s	m	m		16
0B	s	w	m	m	33
1B	ss	w	s	m	100
2B	ss		m	m	71
3B	s	w	w*		32
4B	s	m	m		52
5B	s	ss	m	w	10
6B	s	ss	m*		7
7B	m	w	m	w	27
8B	s	m	m*		11
9B	m	w	m	w	29
0C	s	ss	m	w	4
1C	s	w	m	w	30
2C	m	w	m	w	22
3C	s	m	m	ww	18
4C	s	s	s	w	10

ss = very strong; s = strong; m = medium; w = weak; ww = very weak

* No (001) peak

percentage of residue tend to have the greater concentration of albite. Albite ranges from a trace in samples 5A and 6A up to 60 to 70 per cent of the residue in samples 0C, 5B, and 6B. Indications are that the absolute amount of albite remains fairly constant while the clay and quartz content of the residue varies. This would indicate that the formation of the albite is independent of the amount of volcanic detritus present.

Tuttle and Bowen (1950, p. 76) have shown that there is a difference between the x-ray patterns of high- and low-temperature albite. A study was made of the albite from the K-ben-

TABLE 14.—SPACINGS AND INTENSITIES FOR ALBITE

Amelia albite	I	OH-2 and OH-4C	I
6.38	1	6.4	ww
4.02	3	4.03	w
3.83	2	3.84	ww
3.79	2	3.77	w
3.67	3	3.67	w
3.51	1	3.52	ww
3.38	1	3.38	w
3.20	10	3.21	s
2.96	2	2.96	w
2.93	3	2.93	w
2.86	1	2.86	ww
2.64	1	2.64	ww
2.562	1	2.56	w
2.448	1	2.450	ww
2.412	1		
2.325	1		
2.280	1		
2.190	1		
2.126	1		
2.081	1		
1.893	1	1.890	ww
1.832	1	1.830	ww
1.808	1	1.802	w
1.790	1	1.791	w
1.755	1		
1.723	1	1.727	w
1.676	1		
1.576	1		
1.564	1	1.562	ww
1.535	1	1.529	ww
1.507	1		
1.468	1	1.463	ww
1.437	1	1.431	ww

s = strong; w = weak; ww = very weak

tonite (OH-2) and from the limestone (OH-4C) in an effort to discover any difference between the two. The albite from the K-bentonite is possibly the high-temperature type, and the definitely authigenic variety from the limestone is the low-temperature type.

The albite was concentrated by placing the sample in a mixture of acetone and bromoform and centrifuging at 10,000 rpm. The material which sank and that which floated were examined after each centrifuging, and more acetone or bromoform was added accordingly and the process repeated. In this manner a concentration of approximately 95 per cent albite was obtained.

X-ray spectrometer and film-type patterns show no appreciable difference between the two samples of albite. The *d* values check closely with those listed by Tuttle and Bowen (1950, p. 76) for the low-temperature Amelia albtie (Table 14). This evidence plus the presence of carbonate inclusions in the albite indicates that it is authigenic in both samples.

Heavy minerals.—A detailed study of the heavy separates from three typical samples gave the following results:

(1) OH-1B—This sample lies immediately above K-bentonite OH-2. The nonopaque minerals consist of about 98 per cent apatite, identical in character with the two types found in the K-bentonites. The other 2 per cent consists of idiomorphic zircons. The average grain size of these two minerals appears to be slightly smaller in this sample than in the bentonite. The heavies contain about 30 per cent pyrite which is usually cubic.

(2) OH-2A—This limestone sample (7 feet above OH-2) contains four major nonopaque minerals and one opaque mineral.

BIOTITE: Light pinkish-brown biotite forms 60 per cent of the nonopaques. Some of the grains are hypidiomorphic though most are xenomorphic; they vary from angular to subround and average 0.15–0.20 mm. in size. Many of the grains contain irregular brown inclusions, and rod-shaped inclusions which commonly form angles of 60° with one another are characteristic. These inclusions cover 1–20 per cent of the surface area of the flakes. This biotite is similar to that in the K-bentonites and is presumably characteristic of the volcanic ash.

ZIRCON: Colorless zircons form 30 per cent of the nonopaques. Idiomorphic and hypidiomorphic grains averaging 0.05 mm. and ranging up to 0.115 mm. constitute about 90 per cent of the total. The remaining 10 per cent of zircons are rounded and average about 0.03 mm. These and probably some of the smaller, idiomorphic zircons are apparently detrital, but the majority of the idiomorphic and hypidiomorphic grains presumably have a volcanic source.

TOURMALINE: About 10 per cent of the nonopaques consist of tourmaline grains. These average 0.04 to 0.05 mm. in size and vary from round to subround. They are pleochroic from light to dark gray and from light olive to colorless. The distinct rounding of the grains, plus the fact that none are

found in the K-bentonites examined nor are reported in similar bentonites (Bonine and Honess 1929), indicates a detrital origin.

MUSCOVITE: A few colorless, clear, subround flakes of this mineral can be found.

PYRITE: About 50 per cent of the total heavy separate consists of pyrite cubes, usually in aggregates.

OH-4C—The heavy separate of this sample (21 feet below OH-0) contains about 95 per cent pyrite in the form of cubes and octahedrons joined together in irregular aggregates.

ZIRCON: Colorless zircons constitute 90 per cent of the nonopaques. There are very few zircons in this separate, but the majority of those present were somewhat rounded. The average grain size is 0.04 mm.

TOURMALINE: Approximately 10 per cent of the nonopaques are subround to angular, prismatic tourmalines, which are pleochroic from light green to colorless and average 0.04 mm. in grain size.

A few grains of chlorite and rutile were found.

Significance of heavy minerals.—Assuming that the tourmalines and the rounded zircons are nonvolcanic and that the biotite, idiomorphic zircon, and apatite are volcanic, a rough estimate can be made of the source of the insoluble material in the limestone. (The pyrite is undoubtedly authigenic.)

The idiomorphic apatite and zircon assemblage in the K-bentonites and in the rock approximately 6 inches on either side indicates that most of the material was deposited as volcanic ash possibly directly from the air. Though no direct correlation can be found between the amount of heavy minerals and the total amount of volcanic material, comparison between samples can be made on a relative basis.

OH-2A, 7 feet above the nearest K-bentonite, contains approximately 15 per cent of nonopaque heavy minerals with a nonvolcanic source and 85 per cent which have a volcanic source. OH-4C, 21 feet below the nearest clay bed, contains a much higher percentage of nonvolcanic heavies.

The prismatic shape of most of the tourmalines indicates that they were probably not derived from older sediments but from igneous or metamorphic rocks. The olive, green, gray, and brown suggest that the ultimate source may have been granitic (Krynine, 1946, p. 68).

Clay minerals.—X-ray spectrometer patterns of the limestone residues show that all the samples but one contain considerable clay. Table 13 gives the relative intensities of the x-ray reflections from the 001 planes of the clay material. All but one of the samples give a fairly strong peak equivalent to a nonexpanded 2:1 clay mineral with an 001 spacing of about 10Å. In some of the residues the organic material present prevented good orientation on the x-ray slide, and the 001 reflection was very weak. However, in these cases the 4.47Å *d* value was assumed to be due to the 110 reflection of the 10Å clay. Additional reflections at 14Å and 7Å were obtained from about half the residues indicating a mineral of the chlorite or montmorillonite type. These weak reflections were probably present, but not detected, in all patterns.

Seven representative residues were selected, and x-ray powder photographs were taken of the bulk residue prepared as follows: untreated; cleaned of organic matter; saturated in NH_4OH; saturated in ethylene glycol; and heated to 600°C for 24 hours. Table 15 lists some of the important 00l values of these patterns. Table 16 shows the relative locations of these samples and indicates the relative intensities of the x-ray reflections from the clays present in each sample. The less than one micron fraction was separated from samples 2A, 1B, and 4C. These three samples were also subjected to a variety of treatments, and x-ray photographs were taken (Table 17).

The uncleaned bulk residue of sample 1B (immediately above K-bentonite OH-2) gives an x-ray diffraction pattern with moderately strong line at 14.2Å. A moderate line at 7.0Å and a weak line at 3.50Å indicate the reflections are probably due to chlorite (the 7.0Å 002 reflection does not commonly occur in montmorillonite patterns). In addition a moderately strong band extends from 8.8 to 11.0Å. Such a wide band indicates an interlayering of 10Å and wider layers.

When this bulk residue was heated to 600°C for 24 hours there was a slight shrinkage of the larger spacing from 14.2 to 14.0Å. The 14.0Å line became darker, while the 7.0, 4.70, and 3.50

lines became weaker. This is typical of the chlorite minerals (Brindley and Ali, 1950, p. 26).

The less than one micron fraction of this sample gave a sharp 14Å reflection together with the corresponding 001 series. In addition it gave a fairly sharp line at 10.3Å as compared to the broad band in the x-ray patterns of the bulk material. This fine fraction was saturated with NH_4OH, $CaCl_2$, and ethylene glycol and x-rayed in each case. Table 17 gives the *d* values of the three characteristic x-ray lines for the variously treated clays. These values are in close accord with the mixed layer data for K-bentonite samples. It can therefore be concluded that sample 1B contains a randomly interstratified clay containing approximately 20 per cent of expanded and 80 per cent of non-expanded layers. However, this sample contains a larger proportion of chlorite than sample OH-2. Figure 4 contains a differential thermal curve for OH-1B which is similar to the curves for the K-bentonites and contains a 200°C hydroxyl-endothermal peak caused by the mixed layer 2:1 minerals. However, the 600°C chlorite reaction is much stronger than in the K-bentonite thus confirming the x-ray data.

Sample 2B (immediately below K-bentonite OH-2) is quite similar to 1B. X-ray diffraction patterns of the bulk residue show moderately strong chlorite reflections and moderately strong reflections from an interstratified clay which appears to have approximately the same proportion of expanded and nonexpanded layers as 1B.

X-ray diffraction patterns of the bulk residue of samples 9A, 9B, and 0C are similar. The

TABLE 15.—X-RAY DATA ON TREATED BULK SAMPLES OF LIMESTONE RESIDUES

	001	001	002	003	004
2A-uncleaned	14.0ww	10.1wm	7.0w	4.70w	3.47ww
600°C-24 hrs.	14.0m	9.9m	7.1m	4.70ww	3.50w
+ NH_4OH	14.1w	10.0wm	7.0wm	4.70w	3.50w
E. glycol	14.0m	10.0ms	7.0m	4.70w	3.50w
9A-cleaned	14.0wm	10.0wm	7.2wm	4.70w	3.48ww
1B-uncleaned	14.2m	8.8–11.0m	7.0m	4.70m	3.50w
600°C-24 hrs.	14.0s	10.1mw	7.0w	4.70w	3.50ww
2B-uncleaned	13.9m	8.6–10.8m	6.9s	4.67w	3.50w
600°C-24 hrs.	14.0s	10.1mw	7.1m	?	A
9B-cleaned	14.2w	10.0m	7.1w	4.67w	3.50w
0C-uncleaned	14.0m	10.1wm	7.0m	4.70m	3.48
600°C-24 hrs.	13.8wm	9.9ww	7.0w	?	?
4C-cleaned	14.1ww	10.0m	7.0wm	4.70w	3.48w
600°C-24 hrs.	14.0wm	10.0m	7.0ww	4.70ww	3.48wm

s = strong; m = moderate; w = weak; ww = very weak; A = absent

TABLE 16.—RELATIVE INTENSITIES OF X-RAY PATTERNS OF CLAYS PRESENT IN SELECTED RESIDUES

Samples	Distance from nearest K-bentonite	Mixed layer	Non-expanded 2:1 clay	Chlorite
OH-2A	10 feet		wm	ww
9A	2 feet		wm	wm
1B	0.5 feet	m		m
2B	0.5 feet	m		m
9B	1 feet		m	w
0C	8 feet		wm	m
4C	20 feet		m	ww

m = moderate; w = weak; ww = very weak

patterns for the three samples contain weak to moderate 14Å reflections. The presence of the 002 7Å reflection and the characteristic change in intensity of several of the lines when the residue is heated to 600°C for 24 hours indicate that the reflections are caused by chlorite. In addition, a weak to moderate 10Å reflection is found for all three samples. The other lines associated with the 10Å reflection indicate that the clay is a pure nonexpanded 2:1 clay mineral in contrast to the interstratified clays in the samples described previously.

TABLE 17.—X-RAY DATA ON THE LESS THAN ONE MICRON FRACTION OF THREE LIMESTONE RESIDUES

	OH-1B		OH-2A		OH-4C	
Uncleaned	14.1m	10.3w				
Cleaned	14.0m	10.45s	14.0ww	10.0s	14.0ww	10.0s
NH_4OH	13.9w	10.3ss				
$CaCl_2$	14.2m	10.7–9s				
E. glycol	14.2m	11.4–9.33m				
600°C D.T.A.	13.8s	9.9m				
600°C-24 hrs.	13.6s	10.0m			13.2ww	10.0s
Mixed Layers Lines—OH-1B						
Cleaned	10.45m	5.2w	3.25w			
$CaCl_2$	10.9s	4.92m	3.33w			
NH_4OH	10.3s	5.2ww	3.28w			
E. glycol	11.4-9.3m	5.2b	—			

ss = very strong; s = strong; m = moderate; w = weak; ww = very weak; b = broad

The x-ray diffraction pattern of the bulk residue of sample 2A contains very weak chlorite reflections and weak to moderately intense nonexpanded 2:1 clay mineral reflections. The x-ray pattern of the less than one micron fraction of sample 2A also shows lines of chlorite and a nonexpanded 2:1 clay. However, here the chlorite lines are weaker, and the 2:1 lines are stronger than in the patterns of the bulk sample, indicating that the chlorite is probably coarser-grained than the 2:1 clay.

Sample 4C, in both the bulk and the less than one micron fraction, gives x-ray patterns nearly identical with sample 2A. The nonexpanded 2:1 reflections for 4C are slightly stronger than those for 2A.

Figure 4 shows differential thermal analysis curves of the less than one micron fraction of residues 2A and 4C with the organic matter partially removed. The exothermal reactions between 300° and 450°C are due to organic matter. The large hydroxyl-endothermal reaction extending from 450 to 700°C with a maximum at 600°C is typical though slightly higher than that of the Carboniferous illites. The relatively sharp chlorite peak which occurs at 600°C in sample 1B is apparently masked by the broad nonexpanded clay peak in these two samples. The high-temperature endothermal and exothermal reactions are similar to those of other nonexpanded clays.

Electron micrographs of sample 1B show thin, clear, angular flakes similar to those in the K-bentonite. The electron micrographs of 2A are similar to 1B, but the flakes are less angular and many are slightly rounded.

ORIGIN

The K-bentonite OH-2 is composed largely of randomly interstratified layers of an expanded and nonexpanded dioctahedral 2:1 clay mineral. The layers are in the proportion of 1:4. In addition the clay contains a small amount of chlorite. Samples OH-1B and 2B immediately above and below the K-bentonite bed are composed largely of an interstratified clay identical with that in the K-bentonite. However, these 2 samples contain considerably more chlorite than the K-bentonite.

All other samples examined have a nonexpanded 2:1 clay as their most abundant mineral. In addition nearly all contain varying amounts of chlorite. Samples 9A, 0C, and 9B contain a considerable amount of chlorite but less than samples 1B and 2B. Samples 2A and 4C, which are farther from K-bentonite beds than any of the other samples, contain less chlorite than any other sample except the K-bentonite itself. (Most of the Lower and Upper Ordovician rocks contain a detrital nonexpanded 2:1 clay as the predominant clay mineral.)

The chlorite is rare to absent in the middle of the K-bentonite bed, increases to a maximum concentration at the outer edges of the beds, and in the limestone gradually decreases away from the clay bed.

The petrographic examination showed that the K-bentonite beds and the shaly layers on either side contain only volcanic heavy minerals. The limestones contain both volcanic and nonvolcanic heavy minerals with the volcanic minerals decreasing away from the K-bentonite beds (Fig. 5).

On the basis of the data collected a reasonable theory can be developed to explain the origin of the clay minerals and the duration of the volcanic activity.

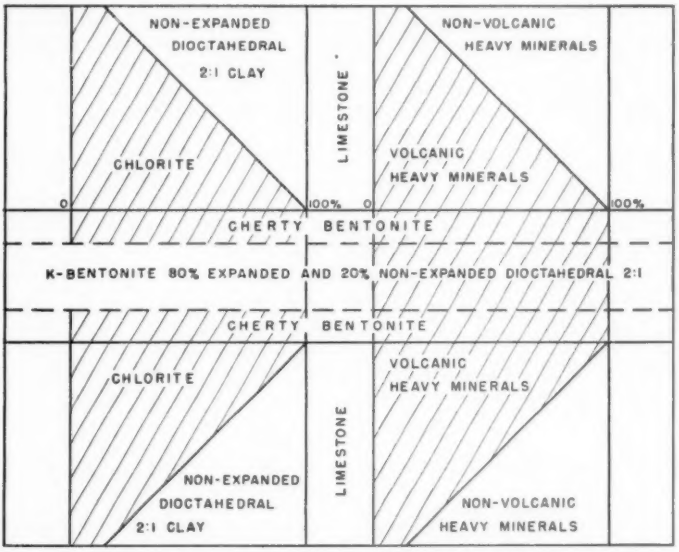

FIGURE 5.—DISTRIBUTION OF HEAVY AND CLAY MINERALS

Chlorite and volcanic heavy minerals increase in relative abundance as the K-bentonite bed is approached. The nonexpanded dioctahedral 2:1 clay and nonvolcanic heavy minerals increase in the opposite direction.

The majority of the latites and trachytes do not contain as much MgO as do the present K-bentonites. Ross and Hendricks (1945, p. 60) concluded that most of the MgO in bentonites, resulting from the alteration of volcanic ash, was derived from the surrounding sea water.

The distribution of the chlorite suggests that the formation of chlorite depended upon the availability of magnesium. This availability is probably enhanced by the presence of calcite ooze. Whereas a plentiful supply of available Mg allowed much of the ash in the outer portion of the clay beds and all the ash in the limestone to alter to chlorite, more restricted circulation of sea water and the absence of calcite inhibited the same sort of alteration within the K-bentonite beds. It is possible that the ash first formed a montmorillonitelike mineral which later altered to chlorite; however, as the Ordovician chlorite is trioctahedral it is un-

likely that it formed from a dioctahedral montmorillonite.

The distribution of clay minerals and heavy minerals indicates that the lime ooze farthest from an ash bed received a small supply of ash which was entirely changed to chlorite. At the same time considerable nonexpanded 2:1 clay was carried into the ooze from the surrounding shore. This resulted in a limestone with a residue composed largely of a nonexpanded 2:1 clay and less chlorite with both volcanic and nonvolcanic detrital heavy minerals (4C).

As volcanic activity increased more ash was deposited along with the nonexpanded clay, and the relative amount of chlorite increased accordingly (9B, 0C, 9A).

The volcanic activity reached a sudden peak and deposited an immense amount of ash in a short time. The amount of nonexpanded clay deposited during this time was negligible. This large amount of ash altered as described (OH-2, 1B, and 2B).

After reaching this peak the volcanic activity tapered off, and the detrital nonexpanded A1 2:1 clay became predominant over the chlorite (2A).

One sample of the middle Ordovician Tioga bentonite was examined; it consists of a mixed-layer clay similar to that found in the Ordovician K-bentonites.

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GRANITIZATION AND HYDROTHERMAL ALTERATION AT BINGHAM, UTAH

BY BRONSON STRINGHAM

ABSTRACT

The rocks exposed in the disseminated copper deposit at Bingham, Utah, consist of folded Pennsylvanian quartzites, limestones, and dolomites within which have been formed granite, actinolite syenite, granite porphyry with associated aplites, and biotite-quartz latite porphyry. Field and petrographic evidence shows that the granite porphyry and biotite-quartz latite porphyry are magmatic. That the granite and actinolite syenite may have been formed by granitization is strongly suggested by: (1) the presence of a microgranitic rock in the granite, recognized as being granitized quartzite, and (2) the replacement of quartzite by feldspar, observable only with the aid of the microscope, that occurs along the contact of the granite. Other chemical, textural, compositional, and other evidences support this hypothesis.

The hydrothermal activity attending the ore deposition took place in seven stages following the formation of the igneous-appearing rocks. Except for the first two, which can be separated on good microscopic evidence, all stages can be differentiated by structural field relations. Stage I saw the formation of kaolinite and illite. During Stage II hydrothermal biotite and sericite were widely developed. In Stage III, chlorite and hydrothermal biotite were formed, but in restricted areas. In Stage IV, quartz and sericite formed over wide areas, while during Stare V quartz was deposited in open fissures without accompanying alteration. Stage VI is the sulfide stage, and Stage VII saw the development of sericite and allophane in open fissures.

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PART I.—GRANITIZATION

INTRODUCTION

The results of the most recent mineralogical study of the copper mine at Bingham, Utah, were published in 1920, and the works of Boutwell (1905, p. 73–390), Keith (1905, p. 29–70), Beeson (1916, p. 2191–2236), and Butler (1920, p. 335–362) were the most extensive. The writer undertook a new study in the summer of 1947; the objective was to investigate the hydrothermal history of the disseminated copper ore deposits. During the course of the work, the possibility that some of the rocks at Bingham were formed by granitization became evident, and after further investigation it was decided to present the full results of the Bingham study in two parts. Part I deals specifically with detailed descriptions of rock units as well as

the granitization and related processes. Part II covers the hydrothermal history related to ore deposition.

FIGURE 1.—INDEX MAP SHOWING THE LOCATION OF BINGHAM

Early workers on the Bingham copper deposit could observe considerably more area than is now exposed since recent dumping of waste has covered several critical areas. On the other hand, since the ore body now has been cut much deeper, less weathered rock and ore are now observable.

The present work has been greatly facilitated by a very favorable geographic coincidence, in that mapping could be done in the field concurrently with laboratory investigations conducted at the University of Utah. This has contributed immeasurably to the quick understanding of the nature of the rocks and has added to the confidence and speed of the mapping.

The Utah Copper open pit mine is situated just south of the town of Bingham, Utah, in the Oquirrh Mountains, a typical basin and range mountain about 20 miles southwest of Salt Lake City (Fig. 1).

The Kennecott Copper Corporation mines the disseminated copper deposit as an open-pit operation employing the level and bank method. The levels carry railroad, power-line, and electric shovel facilities and are separated by approximately 50-foot high banks from which the rock is blasted. The ore contains values in copper, gold, molybdenum, and a little silver.

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LIMITATIONS OF THE STUDY

No geologic map or other such information regarding the mine was received from the operating company, and the data and interpretations presented here were arrived at without any influence other than that gained from the literature. The conclusions are strictly the writer's and do not necessarily represent the views of the operating company. For various reasons, no direct intensive observations were made on the ore body or in areas beyond the open-pit boundaries, and hence no correlation of the disseminated copper deposits with the replacement lead-sliver deposits of the district has been attempted. Since the original plan was to study the hydrothermal alteration and later the granitization, little attention was given to the larger sedimentary or structural features. This present work should be considered as a contribution to only one phase of the general problem. So many new questions constantly arise regarding the geology and mineralogy of the copper deposit proper that it would seem that the present work could still be carried on

for a considerable length of time with profit, and much more work is necessary to complete a full-scale geologic study of the district.

Field Work

Intensive study began in the copper mine in the summer of 1947, and mapping and sampling continued through the field seasons of 1948 and 1949. During the summer months of 1950 and 1951 occasional trips were made to the mine for final checking.

Triangulation points were established on different levels around the periphery of the open pit, and, from these, triangulations were made with the plane table and alidade when accurate locations within the pit were necessary. Most of the work was done, however, with a steel tape and Brunton compass traverse along each level tying into surveyed points at either end of each level. This established fairly accurately, regardless of high-voltage power lines, the positions and distances on each level. The solid rock banks were examined foot by foot, and specimens were collected wherever laboratory examination was thought necessary. In this manner some levels were covered as many as four times, while all levels were traversed at least twice. There was much confusion in identifying rock types during the early stages of the mapping, but considerable confidence was gained after laboratory examination had established the true nature of the rocks. During the entire mapping of the pit no thought was given to granitization, and it was not until the end of the 1949 field season, when a peculiar rock type was definitely recognized as granitized quartzite, that granitization was considered a possible origin for many of the igneous-appearing rocks in the deposit.

Contacts were found only on the banks between the levels in the mine, and extension of these contacts was interpolated along the levels between banks. Solid lines on the map (Pl. 1) indicate contacts observed, and dashed lines represent interpolated positions. Some interpolated contacts were checked when later excavations exposed a new area, and these positions were also placed on the map as solid lines. As is readily apparent, different interpolations of contacts between banks can in many instances produce different map patterns. Some contacts were drawn between banks with a fair degree of certainty, while others are drawn in positions thought to be most reasonable. No major contacts were reinterpolated after the granitization was recognized, and the map stands essentially as it was when first completed. Continued excavations make obsolete most mapped contacts, and in some very active areas contacts are obsolete within a week. Few contacts on cut banks are in exactly the same position as when mapped. The map (Pl. 1) represents geologic conditions during 1947 to 1949 only. Checking in 1950 and 1951 revealed that most of the rock units and contacts are at least in the immediate vicinity of the previous survey. Constant careful mapping of an area being mined at such a rapid rate as this would show some interesting and worthwhile information on the three-dimensional aspect of the deposit.

Almost all rocks collected were thin sectioned, and 803 sections have been studied. Seven large lantern-sized (3¼ by 4 inches) slides were prepared of critical contacts. Many x-ray studies on minerals were made, and in two instances specimens were sent to experts for precise determination. Differential-thermal analysis, universal-stage, and other standard optical equipment was freely used.

The co-ordinates on the map (Pl. 1) are entirely artificial and are placed there only as an aid to the description of the area. Each square is numbered from 1 to 15. These section numbers are used for conveniently locating a specific area.

Chemical Analysis

Ninteen complete and six partial chemical analyses were made on the rocks and minerals at Bingham. Some of these pertain to the hydrothermal problem and are not presented in Part I. However, 13 complete analyses and 2 partial analyses regarded as pertinent to the problem of the origin of the rocks are included.

The quality of the samples available for analysis was poor. Since so much alteration has affected the Bingham rocks probably no entirely fresh specimen was obtained. The difficulties of obtaining a representative sample of Bingham rocks are manifold. The rock breaks into very small fragments due to innumerable post-mineral and pre-mineral frac-

tures, and most of the latter are faced with quartz and/or pyrite which must be scaled off. In some instances, the analyst was supplied with only a very small fragment, and in two instances (Analyses 4 and 5, Table 1) not more than 20 grams could be made available. The large bulk samples which are so highly desirable for analytical purposes were simply not obtainable. The interpretative value of the analyses, therefore, is not so reliable as is usually encountered.

The attempt to calculate a norm from several of the analyses did not work out at all reasonably, and the complications arising because of the extreme abundance of biotite with its variable composition made almost impossible any calculation which would correspond with the actual mineral content of the rock. Measured modes were made in the thin section of each rock analyzed except four which were estimated, and these are all given directly below each analysis. Except for analysis 12—feldspar network, 11—Last Chance stock, and 9—actinolite syenite, all the rocks show a low SiO_2 content and a high MgO, Fe_2O_3, and K_2O content for granitic type rocks. It is believed that the extreme abundance of biotite, being low in SiO_2 and high in MgO, iron, and K_2O, accounts for these anomalies.

General Geologic Setting

The oldest rocks within the area covered in this study are Pennsylvanian quartzites, impure quartzites, limestones, and dolomites which have been folded and locally faulted and within the mine area dip moderately to the northeast apparently without serious displacement. A younger irregular body of quartz monzonite called the Last Chance stock occurs beyond the borders of the mine to the south and west, while within the mine area itself granite, actinolite syenite, and an unusual rock, thought to be granitized quartzite, was formed. Intensive silicate mineral development is found around the periphery of this complex mass in the limestones and impure quartzites, as might be expected along a contact of a granite. The quartzites near the granite have been feldpathized to a moderate degree, visible only with the microscope. All these rocks were intruded first by a large mass of granite porphyry with offshoot dikes, then by a biotite-quartz latite porphyry which appears only in very small dikes and irregular pipes. Extensive hydrothermal activity altered the rocks and culminated in the deposition of the main metalliferous ores.

Sedimentary Rocks

General Statement

All the sedimentary rocks exposed in the pit belong to one formation described and named "Bingham quartzite" by Keith (1905, p. 35), who estimated its total thickness to be 10,000 feet or more. G. H. Girty (Keith, 1905, p. 36) determined its Pennsylvanian age on the basis of two collections of fossils. Gilluly (1932, p. 34), in an area south of Bingham but in the same mountain range, distinguished the upper part of this formation as the Oquirrh Series.

The sedimentary rocks within the pit proper were found to be altered. Since very little work was done during this study on rocks beyond the pit, it is perhaps proper to quote excerpts from the excellent description by Keith (1905, p. 33–35) for the unaltered formation.

> "The Bingham 'quartzite' is composed in the main of quartzites and of sandstones that are more or less silicified. The strata are for the most part white and frequently streaked. . . . In the upper part of the formation are many layers of argillaceous sandstone containing feldspars and mica in fine grains. . . . These sandstones are also slightly calcareous and locally developed into thin limestone beds . . . which are of the same lenticular nature as the limestones that are mapped, and they are so small that it is not practicable to map them or trace them for any considerable distance. . . . Almost any section where the rocks are continuously exposed for 100 yards will show one or more layers of limestone or calcerous sandstone. . . . It is not possible, however, to connect the individual layers of one series with another. Usually the contacts of the limestones and quartzites are sharp and distinct and the transition from one rock type to another takes place within a few inches. In other cases . . . the quartzite passes downward into limestone through several feet of calcareous sandstone and sandy limestone. In still other cases, there is an alteration of limestone and sandstone layers for a few feet. The limestone lenses thin out along the strike and disappear in two ways. Either the closing quartzites approach each other and coalesce . . . or else the quartzites in the strike of the limestones are calcareous. . . . In a few places . . . thin layers of quartz conglomerate appear, but these are fine grained for that class of rock. . . . The uniformity of the grain makes the bedding of the quartzites very difficult to ascertain except where argillaceous, calcareous, or banded beds are present."

If the quartzite contains lime, even as little as 3–4 per cent, local usage classifies it as limestone.

Structure

The general attitude of the sedimentary formations in the pit is quite uniform. Many measurements on quartzite in regions considerably removed from the igneous-looking rocks showed that an approximate strike of N. 70°–85° E. and dip of 25°–40° N. is predominant over the entire area mapped. Near the igneous-looking rocks, alteration and cementation obscure the bedding. Only one sedimentary rock area close to the granite, near the lower center of Section 3, is questionable in regard to a change from the conventional regional dip. The rocks here are extensively altered and probably somewhat faulted, but this faulting is believed to have preceded the formation of the granite. Measurements of the attitude of the beds here were highly questionable, and hence none were placed on the map. A few faults were noted in other areas; their recognition was incidental to the mapping of the limestones. (*See* Section 5.) However, lack of time prevented giving too much attention to this feature since it would not contribute greatly to the major problem. Quartzites and impure quartzites could not be correlated from one level to another. Such a job would require very detailed sampling with accompanying thin sections as well as extremely close surveying control. However, in the mapping of some readily recognizable limestones, it was necessary to call upon faulting to explain their displacement from one level to another (Section 5). Some faults are known to extend into the granite, as extensive slickensided breaks and crush zones are found within it, but most of the major faulting is considered pre-granite. Joints in many attitudes are noted, but no over-all pattern was evident, and an extensive study into this feature was not deemed advisable. The granite rocks and closely adjacent sedimentary rocks at Bingham break up readily under the hammer into very small fragments due to the presence of minute fractures which intricately traverse the rock.

Bingham Quartzite Within the Mine

Pure quartzite.—The grains in the quartzite range from very fine (¼ mm. in diameter) to a somewhat coarse grit (3–5 mm. in diameter).

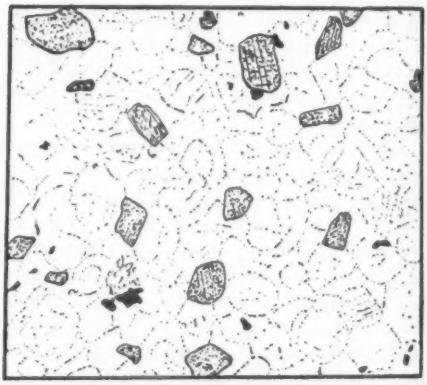


FIGURE 2. ARKOSIC QUARTZITE FROM OUTSIDE THE MINE AREA, NORTHEAST OF BINGHAM

Cloudy grains are orthoclase. Note that ghosts of rounded quartz sand grains are still present. Opaque areas are pyrite. Tracing from photomicrograph. ×30.

No true conglomerate was discovered. Where not stained with limonite, the rock is usually white, gray, or tan. On the upper levels in regions removed from the granite, bedding is plain, but near the granite the bedding is obscure, and the rock is massive and glassy due to strong cementation by silica. Much pyrite is found in veinlets and as small disseminated crystals.

Under the microscope the pure quartzites remote from the granite show some porosity, but the boundaries between grains are interlocking, and ghosts of rounded quartz sand grains are present. In most sections, a little sericite is found interstitially. Rocks near the granite show suturelike grain boundaries and no ghosts of sand grains. Occasional quartzite specimens containing as much as 4 per cent detrital feldspar grains may be found within the pit. These feldspars, usually orthoclase, are scattered in more or less of a random distribution (Fig. 2), and all exhibit smooth, subrounded boundaries.

Impure quartzite.—Many sedimentary rocks differ considerably in composition from the

normal quartzite. Some beds are white and chalklike and much lighter in weight than normal quartzite, suggesting considerable porosity. Some are white to tan, while others range from tinges of green to an intense green. Individual specimens from areas of greenish rocks show blotches, irregular patches, and randomly oriented stringers of deeper greenish material. In many exposures a green streak or veinlet traverses as much as 10 feet through an otherwise massive-looking dense tannish quartzite.

Thin sections of the white impure quartzites show fine interstitial crystals of sericite, tremolite, or, in a few instances, wollastonite. Though generally interstitial the tremolite and wollastonite commonly penetrate quartz grains. Original impurities in these types of rocks are dolomite, calcite, and clay. The pale-green and greenish rocks contain greenish biotite, actinolite, and chlorite. The chlorite is thought to be an unusual variety since its index of refraction, near 1.54, is lower than ordinary chlorites. Quite evidently the greenish colors in the rock are due to different concentrations of chlorite, green biotite, and actinolite. The green stringers that traverse the quartzite may be explained on the basis that the original interstitial impure substances—lime, magnesia, alumina, and iron—were mobile enough during the development of the minerals to migrate through channels and form veinlets or group themselves in clumps and patches.

The impure quartzites may be found on all mine levels where sedimentary rocks occur. However, on the upper levels in regions more remote from the granite, the greenish colors are less prominent, indicating that here the development of these greenish minerals was not pronounced.

Impure limestone.—The impure limestones, which occur in thin beds, are found mainly on the east side of the pit where eight limestone lenses were mapped (Sections 4, 5, 9, 10, 14, 15). One limestone lens was mapped on the west side (Section 2), and two isolated limestone bodies mapped in Sections 2 and 7. Most of these rocks appear dense and greenish. Many specimens contain macroscopic quartz grains, but no carbonate was found. Occasionally some minerals, such as long blades of actinolite or patches of epidote and chlorite, are readily identifiable in hand specimen. The beds range in width from a few feet to as much as 60 feet thick on some levels. The typical lenslike character of the limestones seems to be quite pronounced in some of the beds, and the thickness usually ranges rather markedly along the strike. Introduced pyrite is universally present and locally constitutes as much as 50 per cent of the limestone bed.

Microscopic examination reveals actinolite, chlorite, and occasional epidote and tremolite as the principal minerals of the impure limestone. Quartz grains appear scattered among these minerals. A greenish mineral, which otherwise looks like sericite, is common, as well as garnet which is believed to be near grossularite (index 1.74) in composition. These garnets are rarely seen in hand specimen. One carefully traced bed grades from a rock high in greenish silicates near the granite to a rock lighter in color and containing less silicate minerals, indicating that probably less silica and iron were introduced away from the granitic rocks.

Throughout the granite mass are large and small irregularly shaped inclusions of quartzite and impure quartzite with sharp megascopic borders. Many of the larger inclusions were mapped (Sections 12, 13, and 14 for example), but there are so many in parts of the area that only some could be represented on the map. Bedding on those observed is so obscure that true strike and dip could not be secured with confidence. Petrofabric work on these inclusions as well as the country rock showed no preferred orientation of the quartz grains, and hence no correlation of the inclusion attitude with the dipping country rock could be made using this method.

Dolomitic limestone.—One bed of what is believed to have been a more or less pure dolomitic limestone is found on the southern intermediate levels in Sections 12 and 13. Its mineralogy differs from the other limestone beds. It was identified and mapped principally on the presence of hydrothermal and skarn type minerals, but the outlining of its boundaries seems to be fairly accurate. The bed is judged to be between 30 and 80 feet thick and probably represents a segment of either the Jordan or Commercial limestone member (Keith, 1905, p. 38–41) of the Bingham quartzite. The ap-

pearance of the rock is extremely varied, as different minerals predominate in different areas. The following minerals were found: andradite (index 1.873), grossularite (index 1.74), very fine-grained massive-appearing diopside, actinolite, chlorite, epidote, and much coarse glassy introduced quartz. These minerals occur not in streaks and lenses along bedding, as might be expected, but as irregular patches up to 15 feet or more across. The boundaries between these different minerals are usually very irregular, and veins of one mineral may extend deeply into another. Veinlets of chalcedony, some of which are yellowish, and a few irregular small bodies of magnetite are introduced. In this area occurs the largest irregular veins and bodies of pyrite found in the pit. Fractures which ordinarily cause the rock to break into smaller fragments seem to be absent in this area, apparently because of the silicate minerals, and large blocks generally compose the talus on the bank of the level.

Weathering.—Weathering of the quartzite seems to be confined to the development of limonite, hematite, and some nontronite. On the upper levels near the position of the original ground surface are extensive brown stains. On the east side are areas where limonite and hematite are concentrated enough to color the rock brownish red. Also on the east side, apparently due to channeling of meteoric water, are large streaks of limonite which extend down along the bedding into the white quartzite. On the upper levels, nontronite is extensively developed as a result of weathering of pyrite (Stringham and Taylor, 1950, p. 1065).

Granite

General Statement

The term granite is here used in a broad sense to include several varieties of rock found in the pit. If classified according to Johannsen (1939, p. 141–159) they could be termed orthoclase granite, granite, orthoclase syenite, quartz monzonite, monzonite, syenodiorite, quartz diorite, granodiorite, or diorite. In mapping, all these types were grouped because no structural boundaries separated them, and megascopically they could not be readily differentiated. Seemingly they were all formed by the same process during approximately the same period. All the rocks are rich in biotite, and each rock name probably should be prefixed by the word biotite. The term granite is here used for all the group since normal granite constitutes at least 75 per cent of the whole. The other types vary widely in amount and are found in irregular areas, large and small, randomly scattered over the entire granite exposure.

Early workers classified this rock as a monzonite, quartz monzonite, and monzonite porphyry; it was designated by Butler (1920, p. 351) as the "dark porphyry."

The granite constitutes the largest mass of an igneous-appearing rock within the copper pit and covers about two-thirds of the area. The largest uninterrupted mass, on the lower eastern levels of the mine in Sections 4, 9, 13, and 14, extends high on the south side (Section 12) where it presumably connects through to the granite found in the first 1000 feet of the Niagara Tunnel of the U. S. Mine. A large mass with irregular borders also extends on the upper levels toward the west in Sections 6 and 7. A small isolated area is found on the north intermediate levels (Sections 2 and 3).

Megascopic Features

It is practically impossible to classify the granite megascopically. The quartz, in all hand specimens observed, is in such small units that, except in rare cases, it could not be observed even with a hand lens. Orthoclase and plagioclase are generally both so altered that they appear white to gray and are indistinguishable from each other. Biotite is always recognizable but may be present in such fine flakes that its presence is determined only by its dark color. A granitoid texture characterizes all specimens examined, but four varieties were noted, and a description of each follows.

A medium-grained granitoid texture is found in places over the whole area mapped as granite, and it is by far the most abundant textural type (Pl. 5, fig. 2). In the lower eastern part of the pit where the granite is exposed this texture is very extensive and uniform. The grain sizes range from 1 to 3 mm. in diameter with an average of about 2 mm. The rock is gray. Distinction between orthoclase and plagioclase is

practically impossible except when larger (5 to 25 mm.) pinkish-lavender well-shaped orthoclase crystal are present. In many places these are as much as 20 or 30 feet apart, whereas in some areas they may be within 1–2 feet of each other. The biotite may be euhedral, or it may be in ragged, irregular black or brown patches and may vary in amount from a minimum of about 20 per cent, giving a spotted appearance to the rock, to about 50 per cent when the rock becomes extremely dark or almost black. Some of these dark granites on the intermediate levels of the south side (Section 7) carry plates of biotite up to about half an inch in diameter.

All minerals in rocks with fine granitoid texture are so small that they are not readily differentiated, and the rock takes on a uniform massive appearance. Close examination, however, reveals the ever-present crystals and irregular patches of brownish biotite. The overall color is usually a uniform dirty bluish gray, and the rock somewhat resembles a dense spotted limestone. This fine-grained type is common near the edges of the granite and in small apophyses, but small irregular areas of it may also be found deep within the granite mass. The occasional large pinkish-lavender orthoclase crystals are much rarer in this textural type than in the medium-grained rock.

Coarse granitoid texture is simply a coarser variety of the medium-grained granitoid texture. The individual crystals range from about 2 to 7 mm. across with an average near 4 mm. and therefore offer a much more normal granitic appearance than any other rock in the mine. Despite the larger grain size, biotite is the only definitely identifiable mineral, except for the frequent large pinkish-lavender orthoclase crystals. The quartz, plagioclase, and orthoclase of the main part of the rock cannot be differentiated from one another. This type is found only on the upper western levels of the mine in a fairly restricted area.

In many places the large well-shaped pinkish-lavender orthoclase crystals become so abundant that they make up 30 to 40 per cent of the rock. These are usually set in a groundmass of fine, medium, or coarse granitoid groundmass, and the rock has a porphyritic (or porphyroblastic) granitoid texture. Measurements revealed no preferred orientation or alignment of the crystals. This porphyritic type occurs in scattered irregular small patches at irregular intervals. It is particularly abundant on the southwestern middle levels of the mine.

Near the old ground level the granite has been turned somewhat whitish with brown streaks due to surface weathering. In most cases, however, where weathering is deep, the biotite lightens in color, and a little brownish montmorillonite develops on some of the altered feldspars giving the rock a dull tannish-gray appearance.

Microscopic Features

Except for the size of the crystals, there is only a superficial correlation between the megascopic and microscopic appearance of these granite rock types. Of the 283 thin sections of granite examined, nearly each one is individualistic. However, the mineral species in these varieties are generally quite similar. Thorough preliminary observations revealed six textures which seemed distinctive enough to serve as standards to which all the textures could be referred. However, all combinations of these types exist. The minerals, which are nearly the same in all cases, occur in different proportions, and occasional absence of quartz, orthoclase, or plagioclase, or combinations of these, account for the variety of classifications recognized. Characteristics of each mineral and each texture in the granite type rock are here detailed.

MINERALS: Quartz is anhedral and interstitial. A texture similar to the sieve texture is very common and could be described as a group of interstitial quartz anhedra from 0.1 to 1 mm. across which extinguish simultaneously indicating their optical and crystallographic continuity (Pl. 2, fig. 1). This type is called "sieve" quartz. Quartz may also appear as interstitial dimensional granules .05 to 1 mm. in diameter, but with very ragged outlines and random extinctions. This type is called "granular" quartz. These two types are not to be confused with the "introduced" quartz, which appears in veins, and "alteration" quartz, which is associated with fine hydrothermal biotite and sericite.

Two types of orthoclase occur. Type I crystals are usually small and always euhedral against granular quartz. Of particular note is its anhedral character against itself with most boundaries irregular enough to resemble sutures. Most of this orthoclase is perthitic with a plagioclase (An_{05}). However, all gradations of perthite or antiperthite have been observed from 90 per cent orthoclase and 10 per

cent plagioclase to essentially 90 per cent plagioclase and 10 per cent orthoclase. The grain sizes are less than 1 mm. and average around 0.2 mm in greatest dimension.

Orthoclase of Type II forms the large pinkish-lavender crystals, seen in hand specimen, and are always much larger than the other minerals in the rock. The borders are usually ragged and very thoroughly scalloped, though the crystals as a whole are roughly euhedral. The groundmass minerals quartz, orthoclase, plagioclase, and biotite are in many places incorporated within the large crystal, especially near the margins; the plagioclase inclusions usually are oriented with (010) parallel to the edge of the large phenocrysts. Islands of quartz with scalloped borders are common within the crystals; several islands may have unit extinction. A full chemical analysis and a partial one was made of two of these orthoclase crystals (Table 1, Nos. 13 and 14). The high Na_2O (3.14 and 2.60) is probably largely due to included plagioclase crystals, but some of it may be in solid solution in the orthoclase.

The plagioclase composition ranges from An_{05} to An_{25}; the more albitic plagioclase greatly predominates. Almost all the crystals are euhedral to subhedral except where they constitute cores within orthoclase crystals (Pl. 2, fig. 2). The size ranges from 0.1 to about 2 mm. In most instances albite twinning, although usually present, is somewhat obscured by alteration. Plagioclase cores with orthoclase rims (Type I) of various sizes (0.1–1.5 mm. in diameter) are very common (Pl. 2, fig. 2). The boundaries between the two are extremely irregular, and many are scalloped toward the plagioclase as though the orthoclase were replacing plagioclase. Several islands of plagioclase with unit extinction and optically continuous twinning may remain within orthoclase. Relic or ghostlike alteration patches of sericite having an outline similar to a plagioclase crystal in some places appear wholly within an orthoclase crystal suggesting that this was the position of a former plagioclase crystal which has now been replaced by orthoclase. The proportion of core material to rim material may run from 90 per cent orthoclase to nearly 90 per cent plagioclase.

Biotite, identified by x-ray methods (J. W. Gruner, personal communication), appears in all the granites examined and in some specimens constitutes as much as 60 per cent of the whole. Microscopically much of the biotite forms well-shaped crystals 0.05–1 mm. across. Some of the biotite is prominently scalloped and may poikilitically enclose orthoclase, plagioclase, and quartz. The biotite quite commonly is frayed and ragged due to randomly oriented biotite "foils" around the edges, and some entire crystals, maintaining their original shape, show a mat of fine-grained biotite with identical optical properties; this in reality is a pseudomorph of biotite after itself. In many cases, however, the biotite has altered to secondary biotite accompanied by quartz so that pseudomorphs may range in composition from pure biotite to as much as 5 per cent biotite and 95 per cent quartz (Fig. 8). Fine flaky biotite, as well as sericite, also commonly replaces any of the principal minerals as veinlets or irregular patches. The whole problem of this secondary biotite is more thoroughly described in Part II. A chemical analysis of the alkalies in some mechanically separated coarse biotite was made (Table 1, No. 15). The total alkali content, 10.8 per cent, is near the theoretical figure for biotite, but in this specimen Na_2O seems to be rather high, amounting to .75 per cent.

In all the slides of granite examined, no pyroxene or amphibole was found. However, in 10 of the slides, the outline of the secondary biotite patches simulates the outline of pyroxene and/or amphibole.

In every thin section apatite is an accessory mineral. In some cases it appears in fairly large (0.05 mm.) crystals, but in general crystals are very small and occur interstitially. In almost all slides rutile is present as very small stubby crystals often grouped in clumps. Long needle-like crystals in some slightly altered biotite are also thought to be rutile.

MICROSCOPIC TEXTURES: A true evaluation of the distribution of the different microscopic textures is almost impossible since only the orthoclase prophyritic type can be recognized in the field, and there was no attempt to delineate them in mapping. However, certain generalizations based on the locations of each specimen can be offered. The sutured texture with sieve quartz (2) is by far the most abundant and was found in all parts of the granite area, while the sutured texture with granular quartz (3), next in abundance, was restricted more to the lower eastern part of the pit (Sections 4, 9). Specimens showing the pure sutured texture (1) were encountered only occasionally and in random locations throughout the granite area. The porphyritic and lamprophyric sutured textures appeared principally on the higher levels on the west side of the pit.

The six distinct types of microscopic textures recognized were:

(1) Sutured Texture. Where orthoclase predominates, the extremely irregular boundaries between the orthoclase grains become very striking (Fig. 3). Weinschenk-Johannsen (1916, p. 200) called this texture sutured, and the term fits very

well the texture of much of the Bingham granite. All the orthoclase grains are roughly equidimensional and average around 0.5 mm. across. Biotite and plagioclase may be present between orthoclase grains in this and all the other textures.

(2) Sutured Texture with Sieve Quartz. Orthoclase is euhedral against the quartz but as in texture (1), anhedral and sutured against itself. The presence of sieve quartz is the distinguishing characteristic of this texture.

(3) Sutured Texture with "Granular" Quartz. The quartz appears as small interstitial equidimensional granules between the extremely sutured orthoclase crystals (Fig. 3). The individual quartz grains extinguish independently, as contrasted with the sieve quartz. The boundaries between orthoclase and quartz are irregular in most cases.

(4) Lamprophyric Sutured Texture. This texture may be similar to any of the three textures above, except that most biotite is euhedral giving a texture similar to the type found in some lamprophyres. Biotite usually constitutes 40–50 per cent of the rock.

(5) Porphyritic (or Porphyroblastic) Sutured Texture with Orthoclase of Type II. Large crystals of orthoclase of Type (2) are set in a groundmass of the typical sutured type. In rocks containing abundant large orthoclase crystals, plagioclase is more abundant and slightly basic in composition to An_{25}.

(6) Porphyritic (or Porphyroblastic) Sutured Texture with Plagioclase. In a very few rocks, crystals of plagioclase of about An_{25} are found in a groundmass having a typical sutured texture.

Chemical Analysis

Chemical analyses numbers 1–4, Table 1, are all of granite type rock. Number 1 is the least altered rock found, and this analysis gives the best account of the original composition of much of the granite. Most of the high content of K_2O in this rock is in perthitic orthoclase. Analysis No. 2 was made on the porphyritic variety of granite since this type was always high in plagioclase, and this accounts for the high Na_2O.

Number 3 is of a moderately altered rock of granitoid texture. It is characteristic of a large portion of the granitic rock found particularly in the southern and eastern portions of the granite exposure. The slightly higher Fe_2O_3 and MgO reflect the abundance of the alteration mineral biotite, while the remainder of the analysis approximates analysis Number 1. The specimen for analysis Number 4 was secured about 1 cm. from the granite-granitized quartzite contact (Pl. 5, fig. 4). Here biotite and plagioclase are present in high concentration as indicated by the lower K_2O and high Na_2O and MgO. This rock is near quartz monzonite or granodiorite in composition.

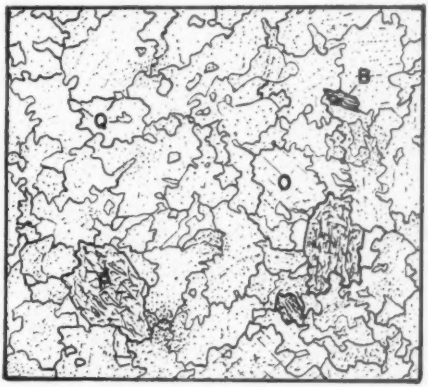


FIGURE 3.—GRANITE SHOWING SUTURED TEXTURE WITH "GRANULAR" QUARTZ

Quartz is in small equidimensional clear grains (Q). Cloudy orthoclase (O) has sutured boundaries. Plagioclase (P) and biotite (B) are also present. Tracing from photomicrograph. ×55.

Contacts and Classification

Contacts between the granite and quartzite, limestone, or granite porphyry are very sharp, and thin sections across the contacts show that it is fairly sharp even on the microscopic scale (Pl. 5, fig. 6). On all contacts observed either in the field or under the microscope, no flow arrangement was found in the minerals of the granite. A petrofabric analysis of the biotite in some of the thin sections across contacts, showed no preferred orientation. However where the granite is next to quartzite, the quartzite exhibits the feldspar network feature, which will be discussed later.

The outline of the contact in places is of particular interest, as in Sections 9 and 10 on the east side of the pit, where beds of limestone and quartzite extend down into the granite mass for two or three levels without any apparent disturbance in the bedding. In Section

TABLE 1.—CHEMICAL ANALYSES OF RELATIVELY UNALTERED BINGHAM ROCKS AND MINERALS

Walter J. Savournin, Analyst
University of Utah

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
SiO_2	68.16	61.73	66.71	65.48	62.28	69.04	67.50	65.39	54.68	61.58	59.42	89.68	65.20*		
Al_2O_3	15.23	18.13	16.21	15.25	16.52	15.77	16.07	14.91	10.80	14.52	15.10	2.78	20.46		
Fe_2O_3	1.29	1.87	2.70	1.88	2.72	1.26	1.70	3.90	5.56	6.54	6.52	1.95			
MgO	2.30	3.06	3.61	4.10	3.48	1.96	2.47	3.27	7.79	4.66	3.93	1.10			
CaO	.62	1.16	.30	1.48	1.21	1.16	.52	.98	11.38	1.41	4.56	.41	.60		
Na_2O	1.24	4.28	.82	3.63	4.29	4.55	2.18	1.71	1.50	2.02	4.04	.57	3.14	2.60	.75
K_2O	8.67	7.74	6.83	4.35	4.87	3.91	7.40	7.00	6.16	5.06	4.03	2.23	9.55	10.25	9.05
TiO_2	.66	.75	.67	.79	.42	.44	.61	.54	.76	.69	.76	.17			
H_2O	1.78	1.59	1.75	2.86	3.91	1.68	1.67	1.84	1.61	3.77	1.01	.20	1.05		
	99.95	100.31	99.60	99.82	99.70	99.77	100.12	99.54	100.24	100.25	99.37	99.09	100.00		
Sp.Gr	2.61	2.54				2.57	2.50	2.63	2.49		2.69				
Modes Measured															
Quartz	25	12	13+	20+	19+	28+	23	19	4	12	9	75			
Ortho. incl. Perthite	62	52	66	28	30	26	56	51	61	34	23	13			
Plagio	5	25	4	32	35	34	8	10	2	22	34	4			
Biotite	7	10	15	18	16	12	12	18	1	32	6	8			
Rutile and Apatite	1	1	2	2	—	—	1	2		—	1	—			
Sphene									2						
Actinolite									11						
Diopside									19						
Augite											12				
Hornblende											15				

Note: All results are calculated to a 105° C. moisture-free and sulfide-free basis. The compensation for the latter is based on the amount of sulfur and copper present calculated to the theoretical formulas of first chalcopyrite then pyrite. The Fe_2O_3 value represents total iron as Fe_2O_3, minus the iron used for pyrite and

Augite											12				
Hornblende											15				

Note: All results are calculated to a 105° C. moisture-free and sulfide-free basis. The compensation for the latter is based on the amount of sulfur and copper present calculated to the theoretical formulas of first chalcopyrite then pyrite. The Fe_2O_3 value represents total iron as Fe_2O_3, minus the iron used for pyrite and chalcopyrite correction. The presence of sulfide prevented analysis for FeO. Any P_2O_5 present is included in the figure for Al_2O_3 but probably does not exceed 0.25%. No calcite was found in any of the rocks chosen for analysis, and hence CO_2 was not determined. The H_2O value was obtained by ignition at 1150°C. corrected for the weight change to be expected from the oxidation of the pyrite and chalcopyrite present.

* by difference.

\+ modes estimated.

− less than 1%.

1. Granite, most unaltered rock available; fine granitoid texture
2. Porphyritic granite, slightly altered; porphyritic granitoid texture
3. Granite, moderately altered; granitoid texture
4. Granite, moderately altered; granitoid; 1 cm. from contact with granitized quartzite; 2 cm. from No. 5
5. Granitized quartzite, slightly altered; 1 cm. from granite contact; 2 cm. from No. 4.
6. Granitized quartzite, slightly altered
7. Granite porphyry, moderately altered
8. Granite porphyry, moderately altered
9. Actinolite syenite, slightly altered
10. Biotite-quartz latite porphyry, moderately altered
11. Quartz monzonite, Last Chance quartz monzonite, very slightly altered
12. Quartzite with feldspar network, slightly altered
13. Orthoclase from porphyritic granite, mechanical separation
14. Orthoclase from porphyritic granite, partial analysis; mechanical separation
15. Biotite from granite, partial analysis; mechanical separation

10 the granite-sediment contacts suggest that the granite extends sill-like into the sediments for a short distance, while in Sections 9 and 14 mapping shows that an interconnecting dikelike network extends deeply into the quartzite in a rather intricate manner. Granite dikes are found near all the margins of the granite and range in thickness from 2 inches to 6 and 8 feet. Many extend very irregularly with unmatched borders, often pinching out or swelling abruptly. No flow structure was found on any of these small dikes, and owing to the homogeneous character of the quartzite no sedimentary rock structures could be traced across dikes with or without offset; thus no evidence was discovered that fully and completely supports either the injection or the replacement origin of these dikes.

In attempting to classify the rock types composing the outcrop mapped as granite, estimates of the relative amounts of minerals were made on all thin sections. Nine modes were determined with the integrating stage, and, according to the Johannsen system, all of these could be classified as a variety of granite (Fig. 4). According to the estimated modes some rocks mapped as granite may have practically no quartz, some may contain very little orthoclase, and others are essentially orthoclase and biotite, giving a range which covers the Johannsen classifications orthoclase syenite, syenite, orthoclase granite, granite, quartz monzonite, granodiorite, quartz diorite, diorite, monzonite, and syenodiorite. No attempt has been made to estimate the areal extent or relative amounts of each type, since they were all determined on microscopic preparations. However, at least 75 per cent of the slides are true granites with perhaps quartz monzonite and monzonite second in abundance. In the porphyritic types containing large amounts of pinkish-lavender orthoclase crystals, the plagioclase content usually increases while the quartz content generally decreases, placing this type generally in the monzonite group.

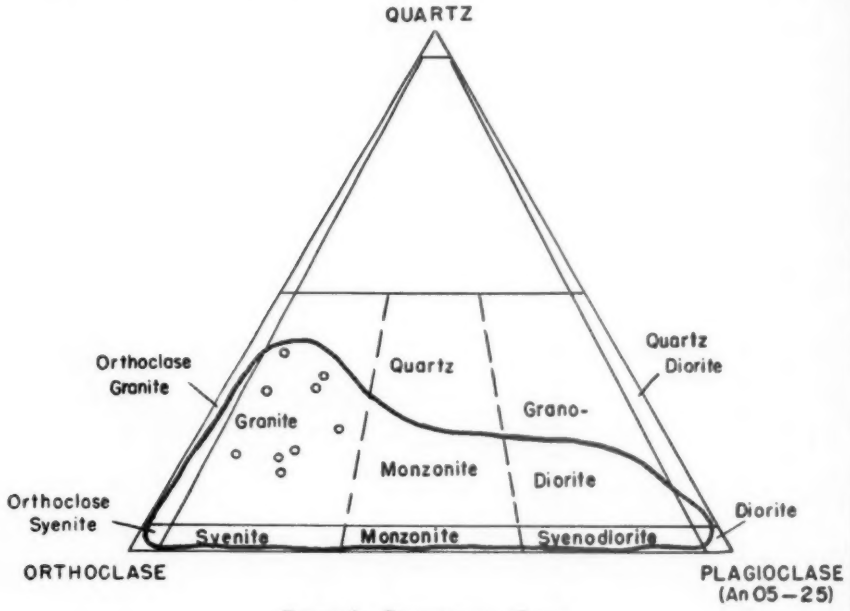


FIGURE 4.—CLASSIFICATION CHART

After Johannsen, showing the mineral composition of the granite. The nine circles represent measured modes and are true granites. Other estimated modes of the granite type rock fall within the area enclosed by the solid heavy line.

Actinolite Syenite

General Statement

The rocks classified as actinolite syenite are mineralogically and texturally similar to those classified as granite but are high in actinolite, epidote, and diopside. This group was mapped as a separate field unit on the presence of the readily recognized actinolite. The term actinolite syenite is used because actinolite and orthoclase are abundant in all these rocks, and quartz is usually absent or present in very small amounts. However, some rocks contain different relative mineral proportions, and other names such as actinolite-orthoclase syenite and actinolite granite are applicable.

The actinolite syenite is confined to areas on the south and southwestern intermediate levels of the pit (Sections 6, 7, 12, and 13) and forms irregular patches and dikelike masses in altered limestone, quartzite, and granite. Similar rocks are also found in the Niagara Tunnel of the U. S. Mine to the south, associated with normal biotite granite.

Megascopic Features

The overall texture of the rock is uniform and granitoid except where large patches of bladed actinolite and massive to granular diopside and epidote occur. The feldspar is gray with a very faint lavender tint and ranges in size from 0.5 to around 1.5 mm. The actinolite is always dark green and is easily recognized; it may amount to as much as 30 per cent of the rock. Biotite can occasionally be seen, but no quartz is discernible. The actinolite, diopside, and epidote, though disseminated throughout the rock in small crystals, may be present also in large (0.5-inch to 3-foot) concentrated patches, streaks, and veinlets.

Microscopic Features

The over-all microscopic features of the syenite are very similar to the granite. Orthoclase is the most prominent and persistent mineral with actinolite, diopside, and epidote very common. Plagioclase (An_5) is rare. The feldspars are roughly equidimensional, but the sizes vary markedly, even within a single thin section.

Quartz, though more often absent than present, is found, as in the granite, interstitially between orthoclase; a number of units extinguish together in the typical sieve texture.

Orthoclase of Type I is also similar to Type I found in the granite. It is roughly equidimensional and often strongly perthitic and with boundaries so

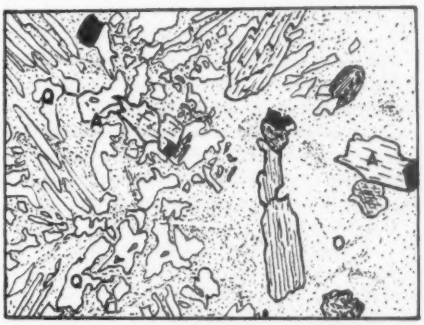


Figure 5.—Actinolite Syenite
The turbid background of the whole illustration is a single optical unit of orthoclase (O). Rods and irregular patches of clear quartz (Q) in left half forms micropegmatite. Actinolite crystals are imbedded poikilitically on right. Note the actinolite crossing micropegmatite boundaries, left center. Traced from photomicrograph. ×50.

irregular they resemble sutures (Pl. 2, fig. 3). Near contacts with quartzite, a structure akin to micropegmatite appears (Fig. 5; Pl. 2, fig. 4), except that the proportions of orthoclase to quartz are usually not in the eutectic proportions; the composition ranges from 90 per cent quartz and 10 per cent orthoclase to 10 per cent quartz and 90 per cent orthoclase. The large pinkish-lavender orthoclase phenocrysts similar to those found in the granite have been observed in only a very few instances.

Plagioclase is not very common, but where found it is usually euhedral with a composition near An_{05}. Plagioclase cores with orthoclase rims, as described under the discussion of granite, are very common.

The biotite in the actinolite syenite is like the biotite described in the granite but is less abundant. Crystals may be euhedral with ragged outline or may poikilitically enclose quartz and orthoclase, and there may be, in rare instances, some fine-grained biotite pseudomorphs after biotite.

Actinolite is always abundant. It is very pale green with X colorless, Y pale bluish green, Z pale bluish green, $\alpha = 1.621$, $\beta = 1.632$, $\gamma = 1.644$, $2V(-) = 72°$, $Z\wedge C = 15°$. Except for a few large crystals, the sizes range from 0.2 to 2.0 mm. The larger crystals average 1–2 cm. in greatest dimension and may be present as single crystals, unoriented clumps of crystals, or arranged in rosette pattern. The actinolite relationship to the other minerals

may be classified in four ways: (1) interstitial but euhedral between orthoclase, plagioclase, biotite, and quartz; (2) poikilitically within crystals of orthoclase (Fig. 5); (3) subhedral crystals crossing boundaries of minerals and boundaries of the micropegmatitic structure (Fig. 5) without disturbing the shape of the actinolite; and (4) irregular patches and veinlets which cut all other minerals.

Epidote is not always present. Under the microscope it is yellowish, $2V(-) = 82°$, $X\Lambda C = 4°$. Crystal sizes range from 0.1 to about 1 mm. It occurs in precisely the same manner as the actinolite: (1) interstitial, (2) poikilitic within orthoclase (3) crossing crystal boundaries of orthoclase, quartz, etc., and (4) irregular patches and veins.

Colorless diopside (α 1.670, β 1.674, γ 1.694, $2V(+) = 52°$, $Z\Lambda C = 38°$) was found in several slides. It is found principally interstitially but it is always euhedral. Occasionally a large patch (3–6 cm.) of very fine-grained diopside was discovered, perhaps representing an undigested dolomite inclusion.

In all slides examined, apatite was present as disseminated crystals of various sizes, from extremely small up to as much as 0.5 mm. in length. Small euhedral crystals of sphene were found in most of the specimens examined, usually disseminated but occasionally grouped in clumps.

Special Feature

Several small crosscutting dikes of actinolite syenite were found in quartzite, having extremely irregular edges, often pinching and swelling within 1–2 inches from a dike 3 inches wide to 0.5 inch wide. The microscope showed that these dikes were rich in micropegmatite of unusual proportions of quartz to orthoclase (Pl. 2, fig. 4) and also rich in actinolite.

Chemical Analysis

Analysis Number 9, Table 1, was made on slightly altered actinolite syenite. SiO_2 and Al_2O_3 are comparatively low here, and Fe_2O_3, MgO, and especially CaO are unusually high as compared to the granite, granite porphyry, and granitized quartzite. The K_2O and Na_2O values are near that of the granite analyses 1 and 3 and granite porphyry analyses 7 and 8. The high Fe_2O_3, MgO, and CaO are due to the abundance of diopside and actinolite. Since this rock contains very little modal plagioclase, the high CaO (11.38) content raises the question of why basic plagioclase is not present in large amounts. This anomaly is discussed in the section on the Origin of the Actinolite Syenite through Granitization Processes.

Contacts and Classification

All contacts of the actinolite syenite with limestone and quartzite are sharp. Small dikes extend into the quartzite, but none was found extending into the limestone. The trace of the contacts themselves are quite irregular, and wall-rock prominences extending into the syenite are common.

The contact with the granite covers 5–10 cm., and two large thin sections (3¼ by 4 inches) covering this contact show that there is no appreciable difference in the texture of the rock, from one type to the other. The only distinction is that the amount of actinolite and epidote of the syenite gradually diminishes to zero, and biotite increases toward the granite. Near the center of the contact, actinolite and epidote are altered partially or completely to aggregates of biotite. This biotite is much more greenish, in contrast to the brown color found in the secondary biotite of the granite. Nowhere along any of the contacts was there any semblance of alignment of crystals.

Only estimates were made of the relative amounts of the minerals in the actinolite syenite. The great majority of specimens contained but a very small amount of quartz and only a little plagioclase. Therefore, the term syenite is applicable in its truest sense to most of these rocks. However, in some slides quartz made up more than 5 per cent of the rock, and these could probably be called quartz syenite, nordmarkite, or perhaps even granite. A few rocks lack plagioclase and quartz and could properly be called orthoclase syenite. Since most of the rocks come under the classification of syenite, however, it seemed advisable to use the term for the group, and, since actinolite is also always present, it should be included in the name.

Granite Porphyry

General Statement

The term granite porphyry includes all the porphyritic rocks found in the pit with the

exception of the porphyritic granite and actinolite syenite, which contain large pink orthoclase crystals, and a biotite-quartz latite porphyry. The relative amount of phenocrysts to groundmass varies rather widely. If a classification based upon whether the phenocrysts constituted more or less than 50 per cent of the rock were employed, nearly one-third of the rocks could be classified as rhyolite porphyry. Compositionally two classifications are possible. In a very large majority of the rocks, orthoclase greatly predominates over plagioclase, resulting in a granitic or rhyolitic type rock. In many rocks, however, the plagioclase-orthoclase ratio is near 1, resulting in a quartz monzonite type rock.

Butler (1920, p. 352) distinguished this rock type with a generalized term "light porphyry"; this term is still in popular use.

About one-third of the total area of igneous-appearing rocks in the pit is granite porphyry. The largest uninterrupted mass is near the northwestern central part of the mine (Sections 3, 8). A large arm extends the length of the pit toward and west where it disappears under dump material (Section 1). Some medium-sized dikes and stringers extending southwestward from the main mass are found in Sections 6 and 7. One small dike here is isolated, and granite surrounds it in outcrop. In Sections 8 and 13 a medium-sized apophysis projects from the main mass toward the south. In the eastern part of the pit the main mass connects with five large dikes and several small plugs (Sections 4, 5, 9, 10, 14, 15). The two large southernmost dikes (Sections 9, 10, 14, 15) crosscut sedimentary beds at a small angle and extend to the southeast beyond the borders of the pit. The southernmost of these was not found to be connected to the main mass in outcrop (Section 9). Near the borders of the large central area are innumerable small aplite dikes which extend into all the different types of adjacent rock. Only a few of these could be shown on the map.

Megascopic Features

All specimens mapped as granite porphyry are distinctly porphyritic with orthoclase or plagioclase and occasionally biotite as phenocrysts. In some cases pheoncrysts may make up 80 per cent of the rock, and unless careful optical observation is made these types may be confused with granite. In some instances, the phenocrysts are so small and so few that the rock appears uniformly dense and may simulate some of the adjacent quartzites (Pl. 2, fig. 5). As a result, thin sections were made to determine the position of several of the contacts. In general, phenocrysts range from 0.5 mm. to 8 mm. in greatest dimension, but in a few instances some are considerably larger. The average, however, range between 1 and 5 mm.

Quartz phenocrysts are found in only a few areas, and most of them are fairly large, glassy equidimensional crystals. Orthoclase is by far the most abundant phenocryst mineral. It is usually white to pinkish, and in many places hydrothermal alteration has made many of the crystals chalklike. In some the rims only are altered. Plagioclase phenocrysts are white when fresh or only slightly altered, but most are intensely altered and then take on an earthy claylike appearance with colors near gray to tan. Biotite phenocrysts are present as brown to black euhedral crystals or in ragged patches of fine-grained hydrothermal material.

The groundmass is very dense and fine-grained, and hence the minerals are not resolvable in hand specimen except where black specks, presumably biotite, are noticed in some rocks having a slightly coarser groundmass. Depending upon the stage of alteration, the groundmass varies from very pale green to white and, where intensely altered, to tan.

Weathering and hydrothermal alteration resulted in two broad general types of changes in the typical granite porphyry.

Type I. Hydrothermal alteration not confined to fractures or other visible channels has produced considerable kaolinite, illite, biotite, and sericite. These minerals weather to montmorillonite which is yellowish brown; in the large central area these brownish-yellow phenocrysts give the rock a distinctly spotty appearance. In hand specimen the montmorillonite is pale green when wet and first collected on a fresh bank, giving the impression that the rock has a high copper content. When dry, however, the brownish yellow of the montmorillonite is seen. Where the rock has been intensely altered and later weathered, the brownish-yellow color

appears in the groundmass, and the whole rock is light tan and chalklike.

Type II. As a result of intense silicification and sericitization, which in this case follows visible channels, hard, dense, quartzose-appearing material traverses the rock in streaks, and the Type I alteration is found interstitially between the streaks. Where channels are so close together that the whole mass is altered, phenocrysts disappear, and the rock resembles dense quartzite with a few small brown patches of biotite.

Phenocrysts predominate over groundmass in the rocks of the large central mass. To the westward, and in the dikes of Section 6 and 7, phenocrysts, larger than average, make up less than 30 per cent of the rock and the groundmass is often slightly greenish. On the eastern side of the pit, in the northernmost dike and the two southernmost dikes, phenocrysts are extremely large but rather scarce. In these dikes large quartz phenocrysts become quite prominent. The groundmass here is also pale greenish. In the small plugs and stringers in Section 5 the phenocrysts are small, less than 0.5 mm. in diameter, and the groundmass is very dense, white, and sometimes chalky due to the intense alteration to sericite.

Microscopic Features

Quartz and orthoclase, which make up at least 95 per cent of the groundmass, are anhedral and equidimensional, presenting a somewhat granular appearance (Pl. 2, figs. 5, 6). The two minerals are generally about the same size. When the phenocrysts are few, the groundmass has an extremely fine texture, but the groundmass grain size grades upward to a maximum of 0.2 mm. in rocks where phenocrysts constitute 80 per cent of the rock. Groundmass plagioclase crystals, which are rather scarce, are usually slightly larger than the quartz-orthoclase aggregate and generally are subhedral. A very few small flakes of euhedral biotite about the same size as the quartz and orthoclase also appear in the groundmass. Apatite, the only accessory, is found in clumps and disseminated crystals.

Quartz phenocrysts are rather large (2–3 mm.) subhedral crystals with extremely irregular borders suggesting resorbtion.

Orthoclase phenocrysts present in all sections examined, make up 30–70 per cent of the phenocrysts. Most of the crystals are subhedral to euhedral but exhibit very ragged borders (Pl. 4, fig. 1). Very commonly quartz and plagioclase inclusions the size of similar grains in the groundmass occur within and along the rims of the orthoclase (Pl. 2, fig. 6).

Misch (1949, p. 384; Pls. 3, 5 of Pt. II) found a somewhat similar condition in the rocks of the Sheku area, China. It is believed that such features of the phenocryst (inclusion of groundmass minerals and irregular borders) may be best explained in either of two ways: (1) The phenocryst could have continued to grow after the groundmass had largely been formed by feeding of late-stage liquid to the phenocryst. (2) The larger phenocryst has less surface energy than the smaller groundmass orthoclase crystals, resulting in an activity gradient where ions making up orthoclase migrate from the smaller crystals to the larger. This results in the growth of the larger crystals at the expense of the small groundmass orthoclase crystals. One effect of this would be ragged borders. Also, the groundmass quartz, and plagioclase originally occupying interstitial positions between similar sized randomly oriented groundmass orthoclase, would now find themselves surrounded poikilitically by orthoclase particularly near the margin. The problem of why some of the plagioclase inclusions become generally oriented with (010) parallel to the phenocryst edge (Pl. 2, fig. 6) has not been satisfactorily solved.

Plagioclase phenocrysts occur in all the porphyritic rocks and usually constitute 5–50 per cent of the phenocryst content. They are generally euhedral to subhedral with fairly straight borders, and their composition ranges between An_{01} and An_{30}. Most of the phenocrysts are so highly altered that the albite twinning is obscured. No plagioclase cores with orthoclase rims similar to those found in granite and actinolite syenite are present in the granite porphyry.

Biotite phenocrysts are not abundant but are always present, composing 10–20 per cent of the phenocryst content. The outlines of the biotite crystals are similar to the outlines of the biotite in the granite and actinolite syenite. Some crystals are euhedral or have a general euhedral outline but with ragged borders, some contain inclusions of groundmass material, and some are changed to secondary biotite (pseudomorphous after biotite). Much biotite is slightly bleached with small needles of rutile arranged within it in the commonly observed trigonal manner.

In hand specimen an alignment of crystals indicating flow is very obscure, but thin sections of a few specimens from near some of the contacts showed a strong flowage alignment parallel to the contacts.

Chemical Analysis

Analyses numbers 7 and 8, Table 1, were made from the least altered granite porphyry specimens obtainable. The SiO_2 and Al_2O_3 contents do not vary much from the granite analyses 1 and 3. The amounts of Fe_2O_3, MgO, CaO, and K_2O are also very similar, and Na_2O is somewhat higher. From these four analyses (1, 3, 7, 8)—two of granite and two of granite porphyry—it seems evident that there is not much difference in the chemical composition of the two types.

Contacts

All observed contacts of the granite porphyry with granite, limestone, or quartzite are "knife edge" and quite regular when carefully traced, so that these contacts were more confidently interpolated from one level to another than were the contacts of the other rocks.

Aplites

Small dikes and stringers related to the granite porphyry are profuse near its borders and extend out into adjacent rock as much as 500 feet. These dikes range in width from 5 mm. to 2 feet and are found in all types of rock of the pit except the granite porphyry itself and biotite-quartz latite porphyry. In hand specimen, the aplites are slightly pinkish and rather fine-grained. They are predominantly quartz and orthoclase with a little plagioclase and a very little biotite. All the minerals are about the same size and are roughly equidimensional with fairly regular borders. Thin sections across very small dikes in granite show extremely sharp borders with a graduation in grain size from the center of the dike to the margin, the smaller crystals being next to the margin (Pl. 3, fig. 1). Many more dikes are present than are indicated on the map, and they are most abundant near the large central main mass of granite

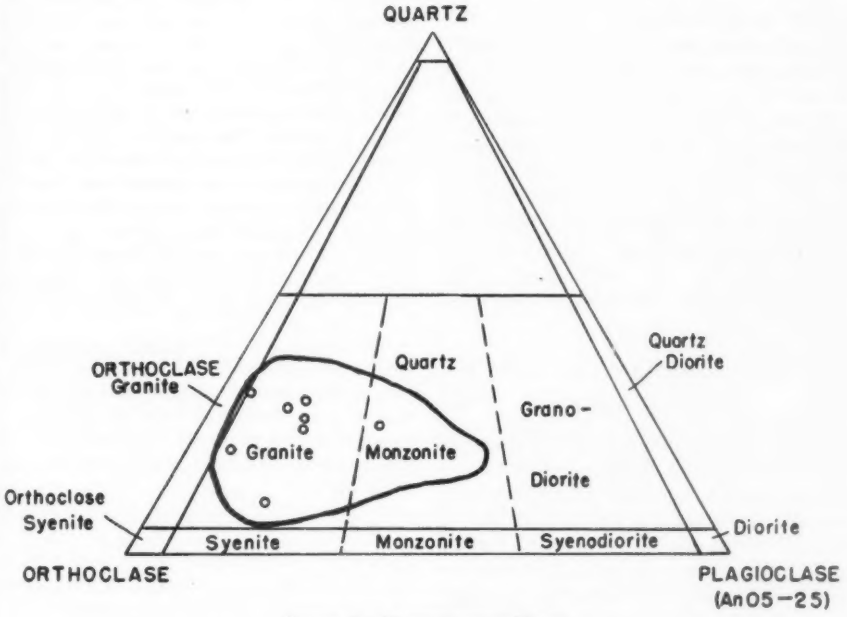


FIGURE 6.—CLASSIFICATION CHART

After Johannsen, showing the mineral composition of the granite porphyry. Of the 8 measured modes indicated, 7 fall in the granite field and one in the quartz monzonite field. All other estimated modes of granite porphyry fall within the area enclosed by the solid line.

porphyry. No offshoot aplite dikes from the dikes or plugs of granite porphyry on the east side of the pit were observed.

Classification

Eight modal measurements were made with integrating stage employing extreme caution to minimize errors due to alteration features. In all but one case these fall within the granite field; the exception is quartz monzonite. An estimation of the modes of all the other specimens showed that in a number of instances plagioclase increases to the extent that the term quartz monzonite porphyry is applicable. These measurements are plotted on a Johannsen diagram, (Fig. 6), and a field is drawn which encompasses all the estimated modal compositions.

Biotte-Quartz Latite Porphyry

General Characteristics

This rock, composed of large white phenocrysts and abundant biotite phenocrysts in a pale-greenish groundmass, is distinctive in hand specimen. It occurs very sparingly in the pit. The rock forms dikes 2–30 feet wide that intrude granite, granite porphyry, and quartzite in the north-central part of the pit. Two dikes were mapped in Sections 2, 3, and 8, but the largest and most accurately outlined mass is in the northwest corner of Section 9 where the boundaries were fairly well located because banks in this part of the pit were moving rapidly at the time of mapping and several return checks could be made.

Megascopic Features

The porphyritic texture is immediately evident in all these rocks. White chalklike feldspar phenocrysts averaging about 3 mm. across are always present, together with very abundant black to brown biotite phenocrysts. The latter are commonly in alignment parallel to the walls of the dike. The groundmass is usually greenish to greenish black; its texture is very fine but in some places is coarse enough so that the grains are resolvable with a lens. Large pink orthoclase crystals, as much as 5 cm. long, occur sparingly.

Microscopic Features

All the smaller euhedral feldspar phenocrysts of the biotite-quartz latite porphyry are so altered that their true nature has been almost obscured. Probably most of them were plagioclase since the orthoclase in the goundmass has not been appreciably altered. The large pink euhedral orthoclase crystals are very slightly perthitic, and some of their borders contain inclusions of groundmass. Quartz phenocrysts with ragged edges are seen only under the microscope and are rather small and comparatively rare. They may be present as single isolated crystals or in groups. Some of the groups of quartz anhedra are elongated as though they had been stretched out due to flowage. Biotite crystals are either euhedral, subhedral, or have very ragged borders, and some are completely changed to biotite pseudomorphs. Most of the biotite shows alignment due to flowage (Pl. 3, fig. 2), and many crystals appear to have been bent.

The groundmass is a granular aggregate of quartz and orthoclase with much biotite mixed with it. Alteration to a greenish biotite and chlorite is always a prominent feature, and no doubt imparts the green color. A little rutile was found as well as considerable apatite; both occur in small, disseminated crystals.

Chemical Analysis

Analysis Number 10, Table 1, is of biotite-quartz latite porphyry. Except for a high content of iron and magnesia accountable for in the profuse biotite, there is not much difference in composition from that of granite, granitized quartzite, and granite porphyry.

Contacts and Classification

All the contacts observed are quite sharp and fairly regular, and alignment of crystals showing flow is quite evident (Pl. 3, fig. 2). The direction of intrusive movement, however, was not determined in the field.

The abundance of euhedral biotite phenocrysts at first suggests that the rock is closely related to the lamprophyres. If the altered feldspar phenocrysts are all plagioclase, and the orthoclase of the groundmass is considered, the rock represents a composition near monzonite or quartz monzonite. Since groundmass always predominates over phenocrysts, the term quartz latite porphyry is probably more applicable. The profuse biotite in the groundmass as well as in the phenocrysts justifies the prefix biotite.

Last Chance Quartz Monzonite

The Last Chance intrusive stock is an igneous-appearing rock outcropping in a large area west, southwest, and south of the copper mine. None of the Last Chance rocks are believed to occur within the mine proper, but previous workers have made important reference to it. The Last Chance quartz monzonite is mineralized only slightly, and no appreciable disseminated copper mineralization has been found within it. Some of the early workers thought that all the igneous rocks in the Bingham District were of the same type and the same age. On his map, Boutwell (1905, Pl. XXXIII) used the same pattern for both the Bingham rocks and the Last Chance rocks. During this study approximately 50 thin sections of Last Chance specimens were observed.

Keith (1905, p. 52) classified the Last Chance rock as monzonite. The rock is medium fine granitoid with grain sizes averaging usually 1–2 mm. in diameter. Average mineral composition ranges as follows: quartz 5–15%, plagioclase (An_{26-40}) 30–40%, orthoclase 20–30%, biotite 5–10%, augite and hornblende 20–30%, and the accessories, magnetite, rutile, and apatite. In augite $\alpha = 1.685$, $\beta = 1.690$, $\gamma = 1.704$, $2V(+) = 60°$ $Z\Lambda C = 47°$, it is faintly pleochroic from colorless to very pale green. The hornblende showed X pale green, Y green, Z green, $Z\Lambda C = 15°$, $2V = 84°$. The texture and mineral composition are much more uniform in the Last Chance rock than in the Bingham granite; the only exception is in some specimens from small dikelike bodies and along the edges of the mass which are porphyritic. Plagioclase is generally euhedral, and orthoclase and quartz interstitial. In several instances hornblende rims augite (Fig. 7). Analysis No. 11, Table 1, shows higher CaO and lower SiO_2 than any of the Bingham rocks other than the actinolite syenite. Evidently the CaO in the Last Chance stock has gone into the making of plagioclase as contrasted with the CaO in the actinolite syenite (Analysis 9), which went mainly into actinolite, epidote, and diopside.

The question as to whether the Bingham rock and the Last Chance rock are the same or not has been a matter of debate for a number of years, but from the evidence the writer believes that the Last Chance rock is distinctly different from Bingham rock.

Feldspar Network

Microscopic Features

Early in this study it was noted that in sections across granite-quartzite contacts the quartzite always contains orthoclase and plagioclase (An_{06} to An_{12}) interstitially, replacing the edges of quartz grains. Not until it was noted that thin sections of specimens taken as much as 300 feet from the granite contact also contain such introduced feldspar was the wide extent and full significance of this feature realized. Because of a late start on this problem and the monumental task of outlining the quartzite areas containing the feldspar through multitudinous thin sectioning, only a beginning on the problem of its areal extent could be attempted.

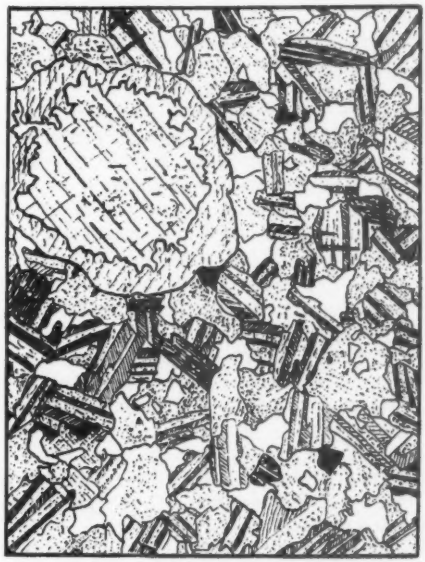


Figure 7.—Last Chance Quartz Monzonite

Large crystal in upper left is augite with a hornblende rim. Twinned plagioclase, orthoclase, a small amount of quartz (clear), and three crystals of biotite are present. Opaque is magnetite. Tracing from photomicrograph. ×40.

Although the hand specimen containing feldspar network looks no different from the normal quartzite, the characteristics observed in thin section are very unusual and striking. In many specimens the quartz grains interlock in the conventional manner, but between many grains are irregular stringers and patches of feldspar (Pl. 4, fig. 2). Only occasionally does feldspar cut quartz anhedra, and this is usually in a pattern suggesting that the quartz was once fractured. The feldspar is otherwise interstitial, interweaving, and connecting one patch with another. Some replacement patches of orthoclase are up to 2 mm. across. The individual feldspar anhedra are usually quite small with sutured borders, but occasionally a single optical unit of feldspar may surround several quartz grains. Stringers may extend from these patches through the interstices of the quartzite to other patches, supporting the belief that all or most of the network may be interconnected. Under crossed nicols most of the sections appear to be normal quartzite; it is only when nicols are uncrossed, the light cut down, and the microscope set slightly out of focus that the network shows up well. Where feldspar is present in large amounts, quartz anhedra may be surrounded. Also, in some cases where orthoclase is abundant, quartz-feldspar boundaries may become fairly straight (Pl. 4, fig. 2), but feldspar-feldspar boundaries are always sutured.

An unusual feature appears in many of these isolated or nearly isolated quartz grains. Their shape is euhedral to subhedral with rhombic shapes predominating and occasionally other short edges appearing (Pl. 4, fig. 2). Bayley (1893, p. 85–96) found a similar condition in the mottled quartzites near the Pigeon Point

PLATE 2.—PHOTOMICROGRAPHS

FIGURE 1.—FINE-GRAINED GRANITE, X NICOLS

Least altered granite observed. Light areas mainly quartz, and darker minerals largely orthoclase. Note a typical patch of sieve quartz (isolated anhedra with unit extinction) within outlined area.

FIGURE 2.—PLAGIOCLASE WITH ORTHOCLASE RIM IN GRANITE, X NICOLS

FIGURE 3.—ACTINOLITE SYENITE, X NICOLS

The predominant mineral is perthitic orthoclase (darker gray) with suturelike boundaries. Very few small grains of quartz (white) present.

FIGURE 4.—ACTINOLITE SYENITE FROM NEAR QUARTZITE CONTACT, X NICOLS

Micropegmatitic structure well developed with quartz (light gray) greatly predominating over orthoclase (dark gray)

FIGURE 5.—GRANITE PORPHYRY (RHYOLITE PORPHYRY) SHOWING VERY FINE TEXTURE, X NICOLS

Phenocrysts are altered feldspar in a groundmass of quartz and orthoclase.

FIGURE 6.—TYPICAL GRANITE PORPHYRY, X NICOLS

Quartz, orthoclase, plagioclase (An_{12-15}), and little biotite compose the groundmass. Phenocrysts are orthoclase. Note the oriented inclusions of plagioclase and the small islands of quartz in the large orthoclase phenocryst (lower center).

PLATE 3.—PHOTOMICROGRAPHS

FIGURE 1.—CONTACT OF APLITE DIKE (UPPER) AND GRANITE (LOWER), UNCROSSED NICOLS

Orthoclase is turbid with low relief. Note sharp contact and gradation in grain size of quartz (clear and higher relief) toward the contact. Opaque is sulfide.

FIGURE 2.—BIOTITE-QUARTZ LATITE PORPHYRY, X NICOLS

Aligned biotite phenocrysts are in a groundmass of quartz, orthoclase, and plagioclase.

FIGURE 3.—PORTION OF FELDSPAR CRYSTAL ALTERED ENTIRELY TO KAOLINITE AND ILLITE (ALTERATION STAGE I), X NICOLS

Dark-gray to black areas kaolinite. Lighter-gray to white flakes illite, roughly oriented in grid pattern.

FIGURE 4.—PLAGIOCLASE CRYSTALS IN GRANITE PORPHYRY PARTIALLY ALTERED TO FLAKY SERICITE (WHITE), X NICOLS

Note grid arrangement of sericite flakes with fresh feldspar between.

FIGURE 5.—VEINLET OF HYDROTHERMAL BIOTITE (B) (ALTERATION STAGE II)

Traversing a felted mass of kaolinite and illite (alteration stage I), X nicols.

FIGURE 6.—LARGE ORTHOCLASE CRYSTAL (O), IN GRANITE (Gr)

Cut by a Stage IV alteration band (IV) composed of quartz and sericite, X nicols.

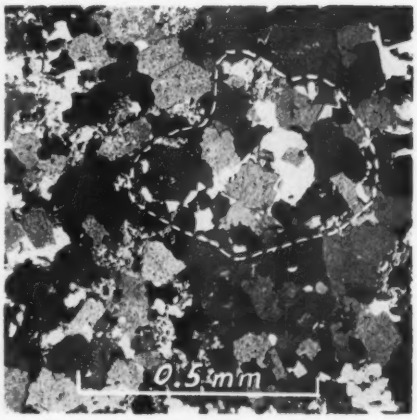
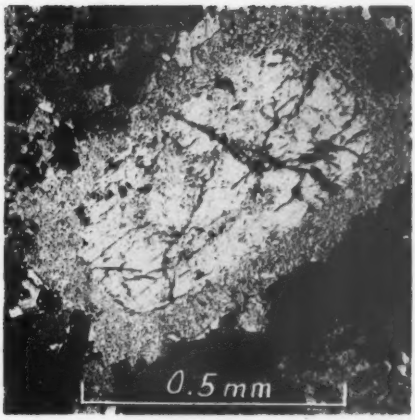


FIGURE 1



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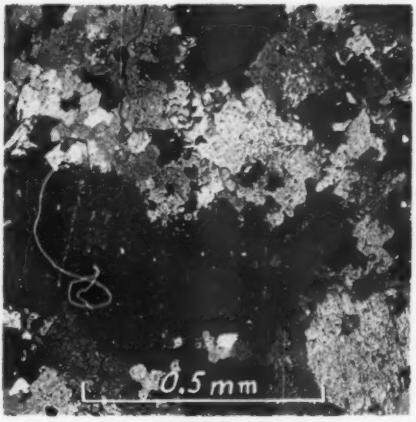

FIGURE 3

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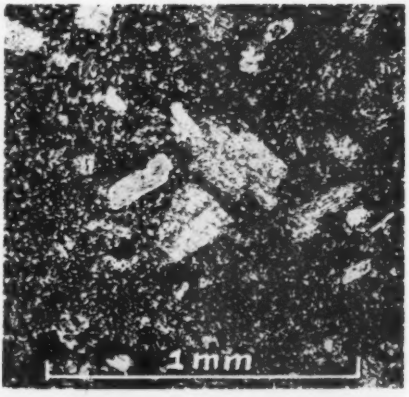

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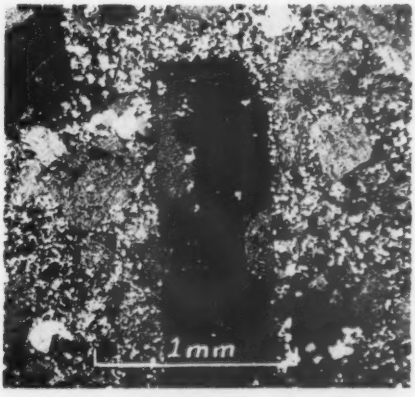

FIGURE 6

PHOTOMICROGRAPHS

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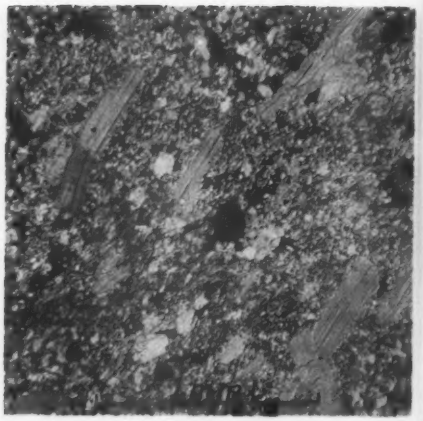

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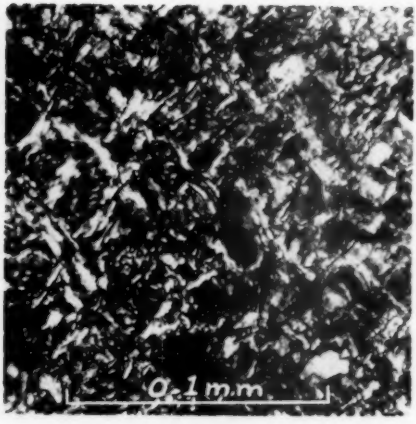

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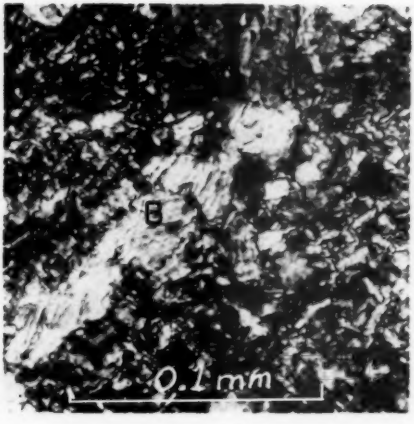

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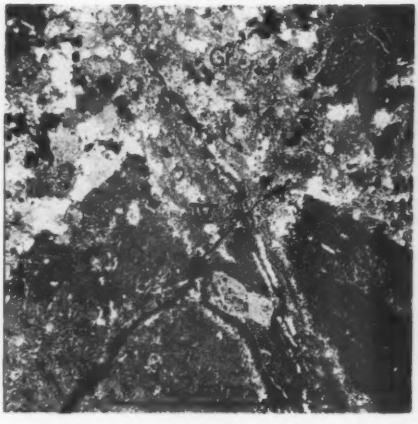

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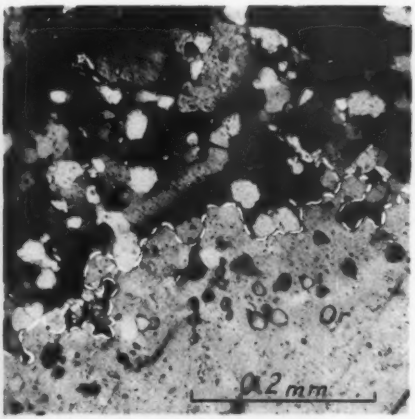

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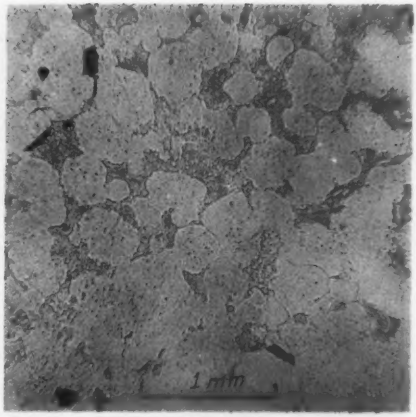

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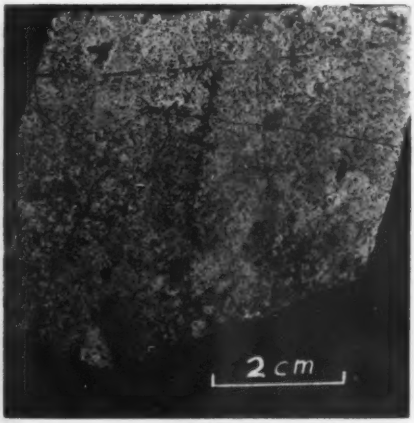

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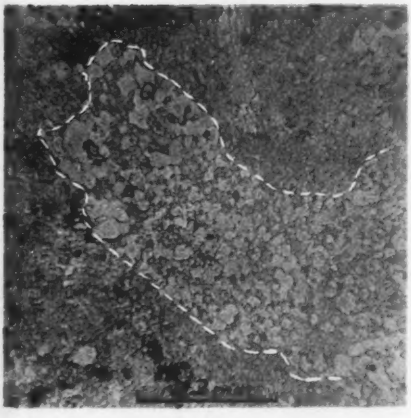

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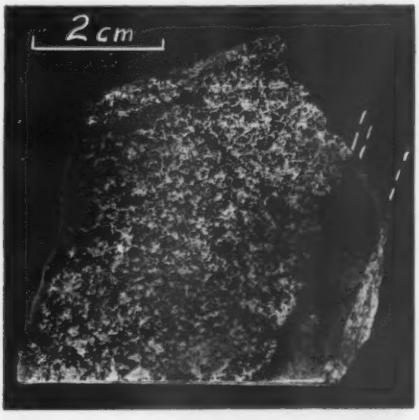

FIGURE 6

PHOTOMICROGRAPHS: GRANITIZED QUARTZITE AND ALTERED GRANITE

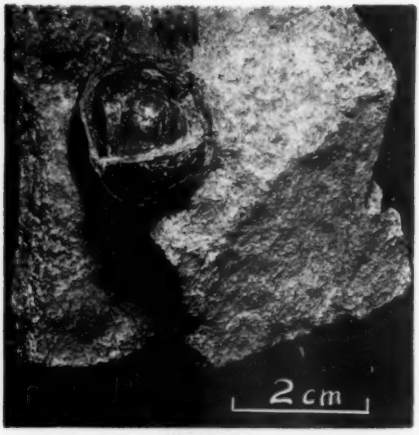

FIGURE 1

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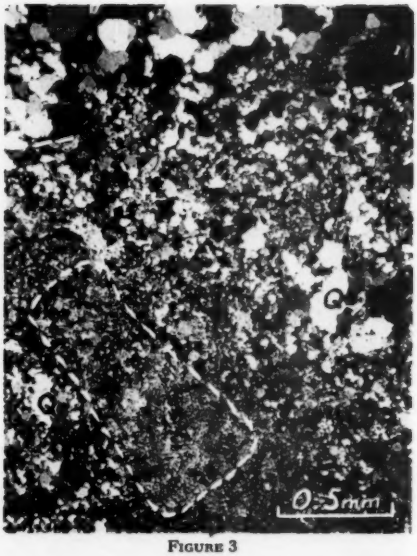

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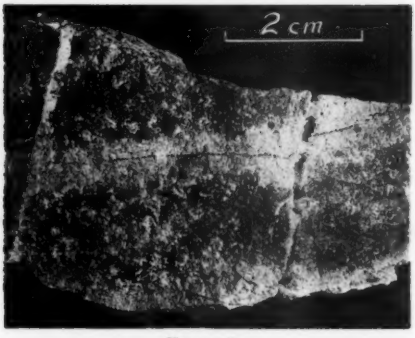

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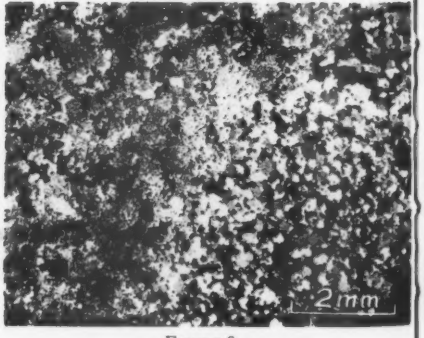

FIGURE 6

ALTERED GRANITE: GRANITE AND GRANITIZED QUARTZITE CONTACT

Sill, Minnesota. Here, however, some of the interstitial material was micropegmatitic. This crystal orientation based on outlines could nearly always be checked by noting the extinction position in relation to the above faces, and often the *c* axis of the quartz could be predicted before confirming with the gypsum plate or quartz wedge. Irregular fragments of some crystallized substances, when allowed to dissolve in certain solvents, are dissolved in particular crystallographic position; after a time the fragment, if extracted from the solvent, exhibits crystal faces (Buckley, 1951, p. 304). Where feldspar replaces along the irregular edges of quartz grains of the quartzite, the crystal structure of the quartz grain controls to some extent where replacement will take place, and irregularities and prominences are preferentially replaced until a fairly flat plane parallel to the rhombohedron or some other crystal face

PLATE 4.—PHOTOMICROGRAPHS: GRANITIZED QUARTZITE AND ALTERED GRANITE

FIGURE 1.—EDGE OF ORTHOCLASE PHENOCRYST (Or) IN GRANITE PORPHYRY, X NICOLS

Groundmass (upper part) composed of granular orthoclase and quartz. Note extremely irregular outline of edge of phenocryst and quartz inclusions near rim entirely surrounded by orthoclase phenocryst.

FIGURE 2.—FELDSPAR NETWORK IN QUARTZITE, UNCROSSED NICOLS

Light-gray high-relief mineral is quartz, and turbid darker-gray mineral is orthoclase. Lower part is quartzite only slightly replaced, but grading upward orthoclase becomes more abundant until some quartz grains are isolated. Orthoclase replaces principally along quartzite grain boundaries. Several replaced quartz boundaries show straight borders some of which prove to be parallel to the more common quartz crystal forms.

FIGURE 3.—PHOTOGRAPH OF POLISHED SPECIMEN OF GRANITIZED QUARTZITE

Shows the very fine texture with the light and dark patches. Veinlets of clear quartz.

FIGURE 4.—GRANITIZED QUARTZITE, UNCROSSED NICOLS

Small, white, high-relief minerals are quartz. Turbid (grayish) low-relief areas are fine aggregates of feldspar (orthoclase + plagioclase (An_{5-10}). Dark minerals are biotite. Near center is a trapezoidal area of feldspar aggregate, probably a former quartzite breccia fragment.

FIGURE 5.—GRANITIZED QUARTZITE, UNCROSSED NICOLS

Quartz grains (Q) imbedded in fine aggregate of feldspar (turbid). Area containing high concentration of quartz is believed to represent an incompletely replaced fragment of quartzite.

FIGURE 6.—GRANITE MODERATELY ALTERED BY HYDROTHERMAL STAGE II (BIOTITE-SERICITE)

To extreme right of specimen may be seen half of a Stage IV alteration band (quartz-sericite) with a dark streak of hydrothermal biotite between it and the main mass of the rock.

PLATE 5.—ALTERED GRANITE: GRANITE AND GRANITIZED QUARTZITE CONTACT

FIGURE 1.—CHLORITE ORBICULE

Showing onionlike structure, developed in granite during alteration Stage III (biotite-chlorite).

FIGURE 2.—POLISHED SPECIMEN OF GRANITE WITH MEDIUM-GRAINED TEXTURE

Darker spots are biotite, lighter areas are either orthoclase, quartz, or plagioclase.

FIGURE 3.—PHOTOMICROGRAPH OF CONTACT BETWEEN QUARTZITE (EXTREME UPPER) AND GRANITIZED QUARTZITE (LOWER), X NICOLS

Lower left is oblong aggregate of relatively pure feldspar. Unreplaced quartzite fragments (Q) appear in the granitized quartzite area.

FIGURE 4.—LARGE THIN SECTION ACROSS CONTACT OF FINE-GRAINED GRANITIZED QUARTZITE (LOWER) AND GRANITE (UPPER), X POLAROIDS

Note that outlined contact is sharp and very irregular.

FIGURE 5.—ALTERED GRANITE

Main portion of rock (dark) is granite altered by Stage II (hydrothermal biotite-sericite). Lighter horizontal streak is a Stage IV (quartz-sericite) band with chalcopyrite and pyrite in the center fracture. Vertical veins are vuggy clear quartz of Stage V, with no adjacent alteration, cutting through the Stage IV band. Small sulphide crystals are present in the vugs as well as disseminated throughout the rock.

FIGURE 6.—PHOTOMICROGRAPH OF GRANITE-QUARTZITE CONTACT, X NICOLS

Contact extends diagonally from upper right to lower center. Quartzite with feldspar network (does not show under X nicols) right. No flow alignment of minerals found in the granite.

has been developed. This face will then be replaced uniformly and the crystal eventually disappears[1].

Much of the feldspar of the network is masked by alteration minerals. In almost all cases one can see a few flakes of fine biotite or sericite as alteration products in the feldspar. Quite often it may be almost completely changed to a solid mat of fine biotite or fine sericite, and very careful observation is necessary to detect the orthoclase that remains. Many quartzite slides, where feldspar network has not been proved, exhibit large interstitial uninterrupted patches of the alteration minerals, but from their sizes and outline it is believed that feldspar network was once present and its original outline represents the present outline of these alteration minerals.

Classification and Distribution

The appearance of the interstitial feldspar in the quartzite gives the over-all impression of a network, and that term is used here.

The determination of the areal extent of the rock containing feldspar network is based on random locations of the thin sections of quartzite examined and thus is of limited reliability. The feldspar network is always found in quartzite and impure quratzite near the granite contact, including all the quartzite and impure quartzite inclusions in the granite. It was also noted in several rocks up to 300 feet from the contact but is more common in specimens located nearer the contact. A few sections of quartzite adjacent to the granite porphyry show no feldspar network, but sampling here is inadequate to conclude that it is absent. Because it cannot be seen in the field, it is not known whether this feature is confined to cross-cutting streaks, certain beds, or is distributed uniformly. The dashed line on the map representing the outer margin implies the probable extent of the orthoclase network. Much further work is necessary to establish the true nature of its distribution. It is surely present, however, and must be reckoned with in any hypothesis regarding the origin of the rocks at Bingham.

[1] Since this paper was written a study of these crystal faces which have developed through replacement has been undertaken, and the report, now in press, will be published in the American Mineralogist.

Chemical Analysis

Analysis number 12, Table 1, of the quartzite with feldspar network, as expected, shows extremely high SiO_2 content. The iron and magnesia is contained in a little biotite which is present as an alteration product. The high K_2O indicates that in this specimen orthoclase predominates over the plagioclase.

Origin

The sedimentary nature of the original quartzite and the evident replacement relationship between the quartz and feldspar indicate that feldspar and small amounts of biotite have been introduced into the quartzite. The possibility was considered that arkosic feldspar grains may be the source of the network feldspar and that it has simply been recrystallized and redistributed. All the true arkoses observed in thin section were definitely low in feldspar content as compared with the feldspar content of much of the feldspar network type rock. Of course if arkosic orthoclase grains were present, they probably were recrystallized and redistributed when feldspar was introduced, since no clastic grains were found in this rock. The amount of feldspar in much of the network exceeds in amount the orthoclase grains found in the most arkosic quartzites. Further, no plagioclase grains were seen in the arkosic types, and plagioclase is abundant in the network. These two features strongly indicate that most of the feldspar of the network has been introduced.

Granitized Quartzite

General Statement

Owing to a circumstance in mapping, the true nature of the peculiar rock classified here as "granitized quartzite" was not fully realized until late in the study. Its origin has now been well established, and this may offer a clue to the origin of the granite and actinolite syenite at Bingham. Earlier workers in the district did not notice this granitized quartzite for apparently two reasons: the writer believes that either there was only a very small outcrop of it at these earlier dates and it was therefore easily missed, or, more likely, it was not exposed at all

but has been uncovered by recent excavation. To the writer's knowledge, no identical rock has previously been described. Details of the rock vary markedly in different specimens, and the mine was revisited many times to try to secure every type.

The granitized quartzite was found near the lower intermediate levels on the southwest side of the pit, specifically in the southern parts of Sections 7 and 8. This area is directly north of some large quartzite bodies which extend into the granite on higher levels (Sections 12 and 13). Contacts mapped are quite accurately located, but interpolation between levels is a matter of judgment, and the true shape of the exposure may not be exactly as drawn. Subsequent location checks, as much as 2 years later, reveal that contacts now (1952) are in somewhat different positions (due to excavation), but the granitized quartzite rock is abundant in this same general area.

Megascopic Features

All the rock in the granitized quartzite area has been so fractured without recementation that rarely can a solid specimen be secured large enough to make a large (3¼ by 4 inches) thin section. To observe the true nature of the rock in the field it must be painstakingly chipped along edges to eliminate the masking effect of innumerable quartz veinlets and expose a fresh surface of the rock. The over-all color is a gray, with white chalklike spots and patches, and brownish spots (Pl. 4, fig. 3). Intensive hydrothermal alteration resulting in abundant biotite has produced a few rather dark rocks. No minerals can be determined megascopically, but in places, particularly where brown spots (presumably biotite) are abundant, the grain size appears slightly coarser. The rock can be best observed on a polished specimen where irregularities in grain size are readily apparent. Streaks and patches of light fine material are seen adjacent to streaks and patches of darker material. These patches may range in size from 0.1 mm. to 1–2 cm., and the borders of most of them appear to be fairly even and straight, giving a superficial appearance of a replaced or strongly recemented brecciated rock. However, the impression is gained that the rock fragments were not fully rotated and separated, but that the fragmentation was more the result of intensive fracturing without much movement.

Unreplaced quartzite fragments or skialiths (Goodspeed, 1948a, p. 520) which range from approximately 1 mm. to several yards in diameter, are very common. Superficially the boundaries of these fragments appear quite sharp, but on the polished specimen, where quartz takes a better polish than the other minerals, it is seen that most of these contacts are gradational. Distributed through the granitized quartzite are many small late quartz veinlets as well as some very small (2 mm.–2 cm. across) discontinuous more coarsely crystalline, irregularly shaped patches and veinlets which have distinct gradational borders with the main rock. The latter are not to be confused with the thicker aplitic dikes which have very sharp borders. Frequently irregular streaks and bands, 2 mm.–3 cm. wide, having a chalcedonic appearance, occur in most of the granitized quartzite rocks examined. Thin-section observation showed that these are due to a late hydrothermal sericitic alteration which spreads more or less symmetrically on either side of small fractures.

Microscopic Features

MINERALS: Quartz grains ranging from .01 to 0.2 mm. in diameter were found in all slides examined. They are euhedral to subhedral and in many cases exhibit at least two straight and presumably rhombohedral edges and often what appears to be a prism edge. High-power magnification revealed that many of the smallest quartz grains have some semblance of crystal outline.

Where orthoclase is adjacent to orthoclase, the borders are always very irregular, but, where it borders quartz, boundaries are fairly straight. The orthoclase grains are slightly larger than the quartz grains, ranging from approximately .02 to 0.4 mm. in greatest dimension. Most of the orthoclase observed was untwinned and only slightly perthitic.

Plagioclase (An_{06}–An_{12}), with the conventional albite twinning (010) and an index near orthoclase or slightly higher, is distinguished from orthoclase principally by the presence of repeated twinning. Most of the plagioclase crystals are slightly larger than the orthoclase anhedra and are more or less euhedral against orthoclase but anhedral against quartz.

There are two types of biotite. One type, believed to be hydrothermal, consists of very fine

flakes scattered somewhat evenly or occasionally concentrated in small clumps within the feldspar areas. The most abundant type, however, is much larger (up to 0.1 mm.) and is believed to have been formed by the same processes that formed the orthoclase and plagioclase. This latter biotite is generally quite ragged in outline. In some cases several ragged and individually isolated crystals have unit extinction; in other cases many crystals may be grouped together with random orientation.

A very small amount of rutile is scattered through the rock, but the crystals are so fine they are detectable only with the highest magnifications. Some biotite, which shows slight bleaching, may contain considerable rutile. Apatite, found throughout the granitized quartzite, forms crystals less than .02 mm. in length. These crystals may at times be concentrated in areas associated with biotite.

The very fine and varied texture usually found in these rocks made it extremely difficult and impractical to make quantitative mineral measurements Since it was estimated that there is a considerable range in relative amounts of minerals in all the slides, it was reasoned that perhaps a measured mode would not give a true picture of the mineral composition of the rock. Estimations indicate the following: quartz 10–20%, orthoclase 30–60%, plagioclase 20–35%, biotite 5–50%, and rutile and apatite combined comprise no more than 1 per cent of the rock.

TEXTURES AND STRUCTURES: The grain sizes of quartz and feldspar vary from .01 to 0.4 mm. The texture at first suggests an evenly distributed aggregate. Various aggregate sizes may be present within 1.5 mm. of each other. Quartz grains are usually detached equidimensional crystals distributed in various concentrations in the feldspar groundmass. Though most of the slides show gradual changes in grain size, many unusual patterns and odd shapes outlined by abrupt grain-size changes are discovered. For instance, quartz, occurring as isolated crystals, may be present in an unusually high concentration in areas covering a small angular patch 0.5 mm. in diameter or perhaps larger (Pl. 4, fig. 5). These areas have fairly straight or regularly curved edges and may be roughly triangular, oblong, or completely irregular. They are considered to represent quartzite fragments partially replaced by feldspar. Quartz, still as detached crystals, is found in straight or regularly curved stringers which often widen into larger angular patches (Pl. 4, fig. 4).

Orthoclase and plagioclase are always interstitial between quartz grains. In no area is plagioclase or orthoclase particularly dominant over the other. Very often an area composed of an aggregate of only feldspar will be bounded by fairly straight or broadly curved sides whose outlines may also resemble triangles, oblongs, or trapezoids (Pl. 4, fig. 4). These are thought to be quartzite fragments completely replaced by feldspar. Most of the hydrothermal biotite occurs in these relatively pure feldspar areas. In addition to these breccialike features, large areas are found in which quartz and felspar are uniformly distributed. In two slides, quartz and orthoclase were intergrown in a micropegmatitic structure, but the relative amounts of quartz and feldspar are not in the eutectic proportions.

Recognizable quartzite fragments are common, irregularly distributed throughout the rock (Pl. 5, fig. 3), and show the feldspar network. Some of the slides indicate that the quartzite apparently was so crushed that fractures traverse individual quartz grains and replacing feldspar has penetrated these cracks.

Three types of "veinlets" traverse the granitized quartzite, in addition to the aplites related to the granite porphyry: (1) discontinuous patches and stringers of slightly coarser orthoclase, plagioclase, and quartz, having gradational borders; (2) megascopally visible veinlets composed of a chalcedonic-looking material which under the microscope is seen to differ from the bulk of the rock only in having slightly more hydrothermal sericite; (3) definite sharp-bordered very small veinlets of colorless quartz.

Contacts

There is a very evident gradational contact between quartzite with feldspar network and granitized quartzite. This gradation is seen in many of the granitized quartzite slides containing recognizable quartzite fragments (Pl. 5, fig. 3). Around the borders of these fragments, which always contain feldspar network, the quartz and feldspar proportions vary with the distance from the contact. Near the quartzite, quartz predominates, while farther from it feldspar becomes more prominent until the usual proportions of quartz to feldspar in granitized quartzite is reached.

Except for the gradational contact with quartzite and the sharp contact with aplitic dikes, granitized quartzite is found in contact only with the granite. In the field this contact appears sharp but irregular, and thin sections across it show that it is very sharp and very irregular, the small grains of the granitized quartzite abutting against the larger grains of granitized quartzite (Pl. 5, fig. 4). In all the

slides studied, no essential mineralogical difference was recognized between the granite and granitized quartzite. Grain size, therefore, is the principal difference between the two, with the granite being the coarser.

Chemical Analysis

Two samples of granitized quartzite were analyzed (Table 1, Nos. 5, 6). Number 5 was taken 1 cm. from the contact with the granite shown in the large thin-section illustration (Pl. 5, fig. 4). The granite specimen, analysis 4, was taken 1 cm. from the same contact. Hence, analyses 4 and 5 represent rock about 2 cm. apart, with a contact between. The range of composition between the two is very slight. Analysis 6 is of a specimen taken some distance within the granitized quartzite area; except for a slightly lower silica content there is not a great deal of difference in the compositions of numbers 5 and 6. Thus the composition of the granitized quartzite seemingly is near that of the high-plagioclase granite (analyses 2 and 4).

Origin and Classification

The original rock from which the granitized quartzite was derived is believed to be either quartzite or impure quartzite. Presumably this rock was intensely fractured, but little rotation of fragments or slipping along fractures occurred. The original rock probably contained much SiO_2 and perhaps a little Fe_2O_3, Al_2O_3, CaO, K_2O, Na_2O, and MgO. From the abundance of the minerals of the granitized quartzite the introduced substances must have included much Al_2O_3, K_2O, and Na_2O with perhaps a little FeO, MgO, and CaO, and a very little P_2O_5 and TiO_2. Of the constituents now in the rock, quartz may be considered residual as well as possibly a small amount of orthoclase of the arkosic quartzite and a very little clay, calcite dolomite, and iron. The new minerals are orthoclase, plagioclase, biotite, rutile, and apatite.

The substances which formed these minerals are believed to have been introduced along the innumerable small fractures and grain boundaries. The fragments were to a certain degree selectively replaced; some were completely changed to feldspar, while others were changed only partially to feldspar and considerable quartz remained. The reason for this selectivity is not known. Much of the rock, where original fracturing was probably not intense, shows the more or less uniform texture of much of the granitized quartzite.

During the later stages of the replacement process, solutions may have become less concentrated in introduced substances causing minerals in localized portions of granitized quartzite to grow larger, resulting in the patches and stringers of coarser minerals.

The term microgranite adequately describes the mineral composition and fine texture of this rock but does not convey the replacement origin or peculiar structure. Since the rock is believed to have been made by a replacement process the name granitized quartzite is considered suitable.

Origin of the Igneous-Appearing Rocks

General Statement

If a conventional intrusive origin for all the igneous-appearing rocks in the Bingham pit—granite, actinolite syenite, granite porphyry with related aplite, and biotite-quartz latite porphyry—were the only hypothesis to be considered, the origin of the feldspar network and granitized quartzite could perhaps be explained as magmatic "fringe effects." Many unusual features of the granite and actinolite syenite, however, need further explanation. Hence, two ideas, origin by igneous activity and by granitization or replacement, are offered here with criteria for each, and the reader may judge the validity of either hypothesis. The writer attempts to present all the arguments on both sides. As the criteria as now set up by proponents of each process do not apply completely, for either case, to the rocks at Bingham, this method of analysis may have merit.

Origin by Igneous Action

Validity of evidence.—Evidences for the sequence of intrusion of the igneous rocks at Bingham, excluding the Last Chance stock, are quite clear except for the granite-actinolite syenite relationship. If the Last Chance rocks are included, there have been either four or five periods of igneous activity, depending upon the origin thought most probable for the actinolite syenite.

Last Chance quartz monzonite.—The position

of the Last Chance rock in the sequence of intrusion has not been definitely established, and this rock does not crop out in the map area. For the purposes of clarity and completeness, its features were discussed previously. Early authors have related it directly to the Bingham rocks. The writer concludes that the Last Chance stock was intruded independently. It is further indicated that the Last Chance magma, of intermediate composition, may have been implaced by stoping and assimilation. Many dikes and offshoots are observed around the periphery of the main intrusive, and in a number of slides taken from these smaller masses a definite alignment of crystals is seen, indicating flowage. The boundary relationships between the minerals in this quartz monzonite indicate that the sequence of crystallization follows essentially the conventional Rosenbusch sequence in that hornblende and augite are nearly always euhedral as is also plagioclase, but the boundary of the latter yields to the boundaries of the dark minerals, and orthoclase and quartz are always interstitial. The fact that hornblende often rims augite may be regarded by some students to mean that magmatic processes have more than likely been operative. The over-all constant uniformity of texture and composition of the quartz monzonite mass and the abundance of a plagioclase of intermediate composition (An_{25-40}), accounting for the higher Na_2O and CaO in the analysis, further strongly suggests that this rock is magmatic. The petrographic and chemical criteria, as interpreted according to the classical petrology, together with the structural relationships, are fairly convincing evidence that the Last Chance quartz monzonite is magmatic.

Actinolite syenite.—The position of the actinolite syenite in the intrusive sequence could not be positively established, but contacts on Plate 1 suggest that dikes of actinolite syenite seem to be cut off by the granite. Structural evidence on the relative ages of the Last Chance rock and the actinolite syenite is not forthcoming since the two have not been found together. However, petrographic features relate the actinolite syenite more closely to the Bingham granite than to the Last Chance quartz monzonite. In this respect, the high content of orthoclase with sutured boundaries, the low amount of albite-oligoclase, and the sieve texture of the quartz are all similar to the Bingham granite and quite dissimilar to the Last Chance rock.

There are several possible modes of origin for the actinolite syenite all of which are based upon a strict magmatic interpretation.

(1) The actinolite syenite may have been intruded early and independently as plugs and dikes in areas within or adjacent to a large bed of dolomitic limestone, the dark minerals actinolite, diopside, and epidote being original orthotectic material.

(2) The actinolite syenite may originally have been a part of a large granite magma. Contamination as a result of assimilation of the dolomitic limestone carrying abundant MgO and CaO could result in a silica-poor granitic-type rock containing enough calcium and magnesium to make actinolite, diopside, and epidote. Serious objection to this idea, however, is that the plagioclase in the actinolite syenite is near albite and not high in An, and, further, it is not at all abundant.

(3) The actinolite syenite may have been originally independently emplaced as a basic gabbroic magma with a chemical composition near the present rock. A gabbroic-type rock may originally have crystallized, but subsequent contact-metamorphic action has recrystallized the rock to actinolite syenite. The writer considers this possibility quite remote.

Metamorphism of the near-by dolomitic limestone may have been accomplished either at the time of the intrusion of the actinolite syenite or at the time of the granite intrusion.

Granite.—The next magmatic stage was the intrusion of the large body of granite which was emplaced in the main by stoping and assimilation of country rock. Since the sedimentary beds are disturbed but very little along the margins of the granite, the intrusion was probably not injected very forcefully. However, as expected along the periphery of a large body of magma, many dikes and small bodies have been intruded into the country rock. Upon examination of the representative dike patterns on the map, one could well imagine that, since many of these dikes interconnect, the stoping process is well illustrated. Inclusions of all sedimentary rock types are common in the granite, and these could be thought of as detached fragments which "floated" downward into the magma from the roof of the magma

"chamber." The crystallization history of the granite could be interpreted to be that of a typical magmatic rock. Biotite is the first and principal dark mineral to develop. The plagioclase crystallized early and was later rimmed by orthoclase. Orthoclase was formed still later with mutual crystal boundaries which are extremely sutured, and quartz crystallized last, occupying interstitial positions. Thus the rock follows the ideal conventional sequence for crystallization from a magma. The wide variations in texture and composition found throughout the granite mass may be explained by local contamination, segregation, or a lack of thorough mixing of the chemical constituents, while the magma was still liquid.

On magmatic reasoning, the feldspar network and granitized quartzite may be explained as a type of fringe effect around a magmatic body where magmatic material to form feldspar and biotite seeps into the country rock and replaces it to varying degrees. The contact metamorphism noted around the edges of the granite could have been the result of the granitic intrusion. The megascopically very sharp borders of the granite with country rock suggest strongly that this was once a solid-liquid contact. Most important of all, the absence of continuous relic country rock structure—in this case bedding—through the magmatic rock further substantiates the intrusive idea. These are felt to be very strong evidences in favor of the magmatic theory.

Granite porphyry.—Following the emplacement of the granite, the large mass of granite porphyry was intruded. This porphyry was probably emplaced chiefly by stoping and assimilating the surrounding granitic and sedimentary rocks. Along the sharp contacts of the granite porphyry are found flow structures which, together with a crosscutting relationship of large granite porphyry dikes, seem to prove that the granite porphyry is distinctly the result of magmatic activity. There are more phenocrysts near the center of the outcrop than toward the edges, which should agree with the ideal crystal distribution of a cooling magma. Since small aplite dikes are found in all rocks except granite porphyry and the later biotite-quartz latite porphyry, the aplite dikes are probably offshoots from the granite porphyry.

Biotite-quartz latite porphyry.—Following the emplacement of the granite porphyry there was a recurrence of magmatic activity resulting in the intrusion of small dikes and irregular pipes of slightly more basic material. Many of the phenocrysts in this porphyry are aligned, showing flow structure, and undoubtedly this rock is truly magmatic.

Origin by Granitization

Actinolite syenite and granite.—The following discussion is more or less confined to features which favor origin by granitization for the granite and the actinolite syenite. The relationship of the Last Chance rock to the granitization process has not yet been fully established, and hence discussion of it is omitted.

It is generally considered that one of the best criteria for the granitization process is an obvious gradation between true igneous-appearing rock and the surrounding unchanged country rock. If this concept is true it is then one of the weakest points favoring granitization of the granite and actinolite syenite at Bingham since borders with sedimentary rocks here appear to be sharp. The feldspar network in the quartzite, though having a sharp juncture with the granite, may be regarded as distinctly a gradational feature between ordinary quartzite and granite. Furthermore, it is a question whether a gradational contact is absolutely necessary for a replacement boundary since many replacement ore bodies exhibit very sharp borders. Also, along the granite contact no flow arrangement of crystals or inclusions has been recognized, either in the field or through a petrofabric analysis of thin sections of granite from areas near contacts.

Perhaps the most convincing criteria favoring granitization is the continuation of relic country rock structures such as bedding, schistosity, etc., from wall rock into or through the igneous-appearing rock. Any such conclusive evidence is lacking in the Bingham granite, but the following could satisfactorily explain its absence. Except for one fairly thick bed of dolomitic limestone, the country rock is largely quartzite with only a few thin limestone lenses. Therefore, sedimentary structures which could be preserved as relics in the granite are practically absent except for the quartzite bedding. If some strong shale beds were present, the

chances for their relic preservation would be fairly good. Since only a few limestone beds are present in the quartzite, possibly this material could be either entirely moved out into the country rock as a sort of feeble basic front where it would be so thoroughly distributed that it would be difficult to detect, or it may have been incorporated and so thoroughly diffused in the replacing material that its identity was lost. Backlund (1946, p. 111) lists rocks in order of their susceptibility to the granitization processes. This is based principally on the chemical dissimilarity of the replaced rocks to the granitizing material. He lists arkose as first and easiest to transform, followed by shale or hornfels, quartzite, limestone, and last, basic rocks. The northwest corner of Section 9 illustrates this, in that quartzite beds have been preferentially changed to granite while altered limestone beds project into the granite for approximately 100 to 200 feet. In some drilling operations an altered limestone was occasionally encountered having granite both directly above and below and its position when projected is consistent with known limestone beds on the surface. This suggests strongly that Backlund's sequence of the susceptibility to granitization of different standard rock types is perhaps operating here and further that the altered limestone extensions occur either as connected unsupported thin beds or isolated undisturbed remnants, and are still in place, not blocks floundering in a magma. Since quartzite bedding is very obscure near the granite, an effort was made to correlate and compare petrofabric information taken from undisturbed quartzite country rock and from isolated quartzite inclusions to see if a common orientation was forthcoming. No convincing preferred quartz diagrams were obtained. The large bed of limestone and adjacent quartzite masses in Sections 12 and 13 are surrounded by granite, and yet their position seems to be fairly conformable to the regional attitude of the country rock. This bed and accompanying group of inclusions could, however, be interpreted as the lower portion of an attached pendant as could also the apparently undisturbed isolated blocks in the lower part of Section 9.

Relationships of wall-rock quartzite to the smaller dikes of granite showed no structural proof of replacement or dilation origin (Goodspeed, 1940, p. 175), but their extreme irregularity, rapid pinching and swelling, and the unmatched wall configuration strongly suggest a replacement origin.

The space problem could be solved better for the granite at Bingham by a replacement process rather than by magmatic intrusion. The structure of the sedimentary beds surrounding the granite show a remarkably uniform continuity except for one small area in the lower right center of Section 3, the disturbance of which is believed to be due to pre-granite faulting. No crushing or brecciation of country rock is evident near the granite, and there seems to have been no major doming or pushing aside of country rock to make room for the magma. Stoping, melting, and other methods of assimilation then are the principal processes to explain the magmatic emplacement.

The concept of a basic front, wherein parts of the material composing the country rock in the area now occupied by the granitized mass have been moved to the adjacent country rock, was first applied (Reynolds, 1947, p. 205) where the removed constituents were somewhat basic. A true basic front is absent at Bingham, but the replaced rocks are quartzite. A silicic front would then be expected to develop, and at Bingham this type of front is thought to be expressed in the silication of the limestone beds and near the granite the introduction of silica into quartzite which has transformed them into rather glassy massive rocks. The true outer limits of this silicic front are difficult to detect in the quartzite, but silication of the limestone proceeds with lesser intensity away from the granite borders.

The extreme suture of the orthoclase boundaries are not what might be expected in a rock which has crystallized from a melt by growth from nuclei, as theoretically in such a case straight boundaries should result (Dan McLachlan, Jr., Personal communication). The sutured texture, then, would appear to be more crystalloblastic than magmatic. Also the extreme variation of mineral distribution even within one section suggests further that the mineral constituents were not uniformly distributed throughout the mass. Some may regard as good granitization evidence the fact that the varia-

tion in grain sizes is in no way related to proximity to the borders of the granite. The ratios of K to Na, which vary markedly in different parts of the Bingham granite, further suggest a nonuniform distribution of the alkalis.

A reasonable explanation of the features of the actinolite syenite is probably one of the best arguments for a replacement origin. The larger outcrops of this rock are close to the large bed of dolomitic limestone in Sections 12 and 13, while the smaller bodies in Section 7 are within or adjacent to quartzite and granite in a region approximately where this limestone projects. All the various textures and mineral assemblages in the actinolite syenite, except the actinolite, epidote, and diopside, are especially similar to the granite. Thus the actinolite syenite could have formed as the result of replacement by the same materials which formed the granite, except that they were contaminated with the CaO and MgO of the dolomitic limestone they were replacing. The capture of SiO_2 originally in the granitizing materials or the replaced rock to make the abundant actinolite, epidote, and diopside, all of which contain high MgO and CaO, might explain the low quartz content of the syenite. Actinolite, epidote, and diopside, associated with orthoclase and a small amount of acidic plagioclase, is a low-temperature mineral assemblage and definitely a peculiar one for an igneous rock. If the rock is magmatic, one wonders why a gabbro-type mineral assemblage did not crystallize, rather than a syenitic type, and why the high CaO was not accounted for in basic plagioclase rather than in the dark minerals. TiO_2 is found as rutile in practically every specimen of granite, yet the TiO_2 of the actinolite syenite is largely tied up in sphene. This could mean that the solutions that formed the granite found a source of CaO in the limestone and formed sphene instead of rutile. Further, the actinolite, epidote, and diopside may be poikiloblastic within orthoclase as well as interstitial between the feldspar, quartz, and biotite, as though they were formed fairly early. Yet these same three minerals are found as euhedral crystals crossing boundaries of quartz, orthoclase, and plagioclase, and also in solid monomineralic patches and veinlets as though they were formed at a late stage as well. This unusual feature is very difficult to explain from a magmatic viewpoint.

In small actinolite syenite dikelets in quartzite where there may have been an oversupply of SiO_2, micropegmatitic structure of unusual orthoclase quartz proportions is abundant. Since this structure is regarded by Drescher-Kaden (1948, p. 205) as a reaction phenomenon, its presence further contributes to the argument for replacement.

Summary.—The above evidences against and in favor of a granitization origin for the granite and actinolite syenite are summarized as follows:

Against Granitization

(1) No prominent continuous relic country rock structures in granite or actinolite syenite.

(2) Sharp megascopic boundaries between granite and wall rock.

In Favor of Granitization

(1) The feldspar network may be considered a transitional zone along the contact.

(2) Limestone extensions and skialiths in granite, aligned with limestone beds of undisturbed country rock. Obscure quartzite bedding would likely not be preserved as a relic in replaced granite.

(3) Periphery dikes very irregular with prominent pinch and swell features. Wall-rock borders largely unmatched.

(4) No flow structures along contacts, even of smallest dikes.

(5) Silicic front present near contacts (glassy quartzite with obliterated bedding, limestone silicated).

(6) Wide variation of textures and rock types in granite.

(7) No miarolitic cavities.

(8) Grain-size variations bear no relation to contacts.

(9) Crystalloblastic nature of microtextures.

(10) Fairly wide variation in K-Na ratios in different areas.

(11) High CaO rock (actinolite syenite) only near large limestone skialith. (a) CaO makes actinolite, diopside, and epidote rather than

plagioclase. (b) Actinolite, diopside, and epidote formed late as well as early.

(12) Strong micropegmatitic structures of unusual orthoclase-quartz proportions suggest replacement.

Granitization history.—If granitization is regarded as a possible origin for the granite and actinolite syenite, the subsequent intrusion of the granite porphyry with associated aplite, and biotite-quartz latite porphyry must be explained.

All the igneous-looking rocks of the area may be explained by two granitization hypotheses: (1) pure granitization, and (2) granitization related to magmatic action.

According to (1) granite-forming materials were transported upward from a source at depth, possibly by hot solutions or vapor rich in water. The quartzite and limestone areas now occupied by the granite were slowly replaced, and the granite and actinolite syenite were formed. Part of this process was the development of the feldspar network and silicic front in the country rock ahead of the advancing granitization front. In one place the replacement was perhaps only partial, and the granitized quartzite rock remaining represents an intermediate stage between the formation of the feldspar network and the normal Bingham granite.

With continued heat, temperatures of the rock in some areas were elevated until the granite became mobile through partial or complete melting forming a neomagma (Goodspeed, 1948b, p. 72). This gave rise to the large mass of granite porphyry which, around its borders, moved into cracks and openings in the granite and quartzite and formed dikes. Differentiation was taking place at depth in this molten mass, and after a portion of the granite porphyry had solidified the slightly more basic biotite-quartz latite porphyry was intruded. The porphyritic nature of the granite porphyry could have developed in two ways. If only a portion of the rock were melted and some remnants of the larger minerals were still solid, upon cooling they would form nuclei for recrystallization causing them to grow to extraordinary sizes and the interstitial melted material would crystallize as groundmass. The alternative hypothesis would be that the granite was completely melted and the phenocrysts developed in the conventional magmatic manner. In either case the question could not be convincingly settled since, under equilibrium conditions, there would be no difference between two magmas of duplicate composition, one of which had been cooled from a liquid until there was 10 per cent of the whole present as crystals in the liquid, and the other which was formed by gradually heating and melting the solid rock until only 10 per cent remained as crystals in the liquid. The textures, subsequent crystallization sequence, etc., would be identical. The more uniform chemistry and mineralogy of the granite porphyry could perhaps be explained by the stirring action which would accompany movement and the chance for more thorough diffusion of chemical constituents afforded by the fusion.

Hyphothesis (2) involves granitization resulting from magmatic action. From a magma at depth would develop granite, actinolite syenite, granitized quartzite, and feldspar network as above, but the whole action would be a wide fringe effect around the upper periphery of the magma. Later, the parent magma could have stoped or assimilated its way into and beyond the granitized rock forming the granite porphyry and later the biotite-quartz latite porphyry.

Evidence for either hypothesis is lacking, and the reader must fall back on his beliefs for his selection.

Part II.—Hydrothermal Alteration

Introduction

Following the formation of the igneous-appearing rocks in the Bingham deposit, secondary processes in the form of hydrothermal alteration and mineral introduction became operative. The alteration and mineralization took place in seven distinct stages, all of which can be clearly delineated. The separation and paragenetic sequence of all but the first two are

established on recognizable structural crosscutting field relationships, while the first two are separated on the basis of crosscutting structure seen under the microscope. During the first four stages the original minerals in the rocks were altered, while the last three stages represent mineralization through introduction essentially without accompanying alteration. In Stage I, rock in a restricted area near the north-central part of the pit was altered with kaolinite and illite forming principally at the expense of the feldspars. Stage II saw the extensive development of sericite and hydrothermal biotite in essentially all the rocks within the mine. The alteration minerals of both these stages are disseminated throughout the rocks and do not follow fractures or other recognizable channels. The amount or intensity of alteration varies widely. These effects may be thought of as having been accomplished by solutions which more or less soaked uniformly through the rock. During Stage III, profuse chlorite and hydrothermal biotite were formed within rather restricted areas of granite and are found in wide bands, the positions of which were obviously controlled by fractures. In Stage IV, quartz and sericite were formed on either side of fractures and appear as bands and streaks in all the rocks in the area. Quartz, the only mineral formed in Stage V, is found in sharp bordered veins with no accompanying alteration. In Stage VI disseminated sulphides were developed in all the rocks within the mine. No alteration in the strict sense accompanied this mineralization stage. Sericite and allophane of Stage VII were introduced into cracks and vugs in somewhat restricted areas.

Chemical Analysis

The chemical analyses of the altered rock presented in Table 2 were made by precisely the same analytical methods as were used with the unaltered rock samples. These figures represent results obtained from rocks observed to be the most intensely altered specimens of their type. The same caution in evaluation of the analytical figures is imposed here as for the unaltered rocks, in that large representative bulk samples were not obtainable. With many analyses of altered rocks, authors have been able to present analytical data that showed progressive alteration intensity (Lovering *et al.*, 1949, p. 41–60). The essential absence of fresh rock, the rather wide compositional variation of rock types, and the complexity and overlapping of the numerous alteration stages at Bingham prevented use of this valuable technique.

Table 2.—Chemical Analyses of Altered Rocks from Bingham

Walter J. Savournin, Analyst
University of Utah

	1	2	3	4	5
SiO_2	68.19	63.06	61.60	74.19	73.61
Al_2O_3	14.67	15.83	14.29	14.41	15.76
Fe_2O_3	1.93	3.74	5.10	.06	1.08
MgO	3.26	4.12	7.44	.77	.89
CaO	.34	1.48	.74	.45	.45
Na_2O	1.96	4.05	1.13	1.00	.78
K_2O	6.25	4.68	5.69	5.60	3.58
TiO_2	.63	.72	.99	.58	.42
H_2O	2.06	1.76	2.49	2.17	2.65
Total	99.29	99.44	99.47	99.23	99.22
Sp.Gr	2.54	2.61	2.70	2.46	

1. Altered granite porphyry. Stage I, Illite-kaolinite
2. Altered granite. Stage II, Hydrothermal biotite-sericite
3. Altered granite. Stage III, Hydrothermal biotite-chlorite
4. Altered granite porphyry. Stage IV, Quartz-sericite
5. Altered granite. Stage IV, Quartz-sericite

Hydrothermal Stage I—Kaolinite-Illite

General Statement

The earliest hydrothermal activity following the formation of the igneous-appearing rocks resulted in the development of kaolinite and "a poorly organized mica possibly in the illite[2] group" (W. F. Bradley, Personal communication). This alteration effect is found only in the granite, granite porphyry, and biotite-quartz latite porphyry. The very small sizes (.001–.003 mm.) of these minerals and the small patches (.05–2.0 mm.) in which they

[2] The name illite will be used hereafter to designate this mineral.

TABLE 3.—HYDROTHERMAL STAGES AT BINGHAM

	Alteration Stages					Mineralization Stages	
	Disseminated		Channels			Disseminated	Channels
	Stage I	Stage II	Stage III	Stage IV	Stage V	Stage VI	Stage VII
Minerals developed	Illite Kaolinite	Hydrothermal biotite Sericite	Hydrothermal biotite Chlorite	Quartz Sericite	Quartz	Chalcopyrite Bornite Digenite Enargite Molybdenite Pyrite Gold Silver	Sericite Allophane
Paragenetic evidence	Petrographic	Field	Field	Field	Field	Field	Field
Introduced materials		Fe Mg	Fe Mg (much)	Si	Si	Cu Mo Fe S As (little) Au Ag	K Al SiO_2 H_2O
Rocks affected	Granite Granite porphyry Aplite Biotite-Quartz Latite Porphyry	All rocks	Granite	All rocks (except sediments)	All rocks	All rocks	All rocks

Time ⟶

occur make it impossible positively to distinguish this type of alteration in the hand specimen, except in a superficial way where a brown chalky appearance is present caused by abundant brownish montmorillonite which has developed through weathering of the alteration minerals. In some granite porphyry rocks intense alteration is represented on a fresh-cut bank where altered phenocrysts may break out readily from the altered groundmass. However, in all cases a thin section was necessary to distinguish this alteration type positively.

Microscopic Appearance

Under the microscope the kaolinite and illite are intimately mixed in almost all instances. Only in a few cases and under high magnification, was one mineral observed without the other. Optically the kaolinite is readily identified by a low birefringence (.007) (Pl. 3, fig. 3). The individual crystals of some aggregates of kaolinite are so small that they often appear isotropic. In most of the kaolinite aggregates, the individual crystal units are roughly equidimensional, but some typical wormlike structures with transverse flakes are present (Ross and Kerr, 1931, Pl. 41). The relief is slightly higher than the relief of quartz.

Illite is readily distinguished from kaolinite under crossed nicols, by its slightly higher birerefringence, with colors in first-order yellows. It has a distinctly flaky habit with individual flakes length slow and a relief slightly higher than kaolinite, the lowest index being about 1.569. Almost all the flakes of illite are intimately mixed with kaolinite, and many are arranged or oriented in a somewhat crude pattern with the angle of the grid being about 90° (Pl. 3, fig. 3). No large patches of pure illite were discovered, and hence it was not possible to work out its precise optical properties. Because of the intimate mixture of the two minerals, it was impossible to obtain a pure enough sample of either one for chemical analysis. Positive identification of the two minerals was made by W. F. Bradley of the Illinois State Geological Survey who x-rayed some completely altered phenocrysts of the granite porphyry. He states that, from his powder pattern, he would judge the material to be "about 50 per cent kaolinite and 50 per cent a poorly organized mica, possibly in the illite group."

The two alteration minerals have formed at the expense of all the primary minerals in the rocks except apatite and rutile. Plagioclase is usually most nearly completely altered, but orthoclase is quite often altered, and quartz is affected around its borders only in the most intensely altered rocks. Biotite is altered only rarely and then also very lightly around its borders.

Wherever a Stage I alteration patch is found it is composed of practically 100 per cent alteration minerals. In only rare cases are these alteration minerals scattered with fresh unaltered mineral between. Where alteration has been very slight, the kaolinite-illite patches may be quite irregular and occupy only a portion of a crystal. With more intensity, a whole feldspar crystal may be affected, and the outline of the alteration patch is that of the crystal.

Distribution

The first alteration stage is most intense near the north-central portion of the pit in Sections 2, 3, 4, 7, 8, and 9 (Pl. 1). The outer boundary of this area was determined by observing thin sections from well-located samples. Within the boundary are rocks altered as much as 75 per cent, but most of them have been altered much less. An arbitrary figure of 15 per cent was selected to determine the outer boundary of the alteration area which appears on Plate 1. Thus, if the rock showed less than 15 per cent kaolinite and illite, it was not included in the altered area. The boundary thus drawn is therefore not absolutely accurate, but its position should be a fairly good approximation of the outer limits of the alteration.

Character of Altering Solutions

Since kaolinite forms principally in an acid environment below 350°C., its presence as an alteration product here would definitely place the pH of the altering solutions of Stage I in the acid category and at temperatures below 350°C. The field of formation of illite is not so well known, but the intimate association between illite and kaolinite would suggest that they were formed at the same time.

Analyses No. 1, Table 2, was made of a specimen of granite porphyry composed of about 60 per cent Stage I alteration minerals. Comparison of these figures with the analyses of moderately altered granite porphyry, Nos. 7 and 8, Table 1, shows that there is very little significant chemical difference between unaltered rock and altered rock of Stage I type. The K_2O and CaO are slightly lower than the unaltered rock, while SiO_2 is slightly higher. Thus it may be stipulated that the altering solutions of the first stage only effected mineral changes, and very little material was added or taken away from the rock.

The alteration follows no obvious channels which can be noted in the field or thin section. The solutions evidently penetrated the rock along mineral boundaries or through microscopic fractures. Since there is no pattern of the latter, it is believed that the solutions resulting in the first stage of alteration penetrated the rock as a "soaking action."

Paragenesis

Stage I is definitely the earliest alteration stage and occurred after the primary minerals in the rocks were formed. The alteration minerals of Stage I are all cut by veinlets or replaced by all other types of secondary mineralization in the pit.

Hydrothermal Stage II—Hydrothermal Biotite-Sericite

General Statement

The alteration produced during the second stage of hydrothermal activity resulted in the profuse development of fine flaky biotite (here called hydrothermal biotite) and sericite. These alteration minerals are disseminated throughout the rock and are rarely found in concentrated clumps, patches, or veinlets. Both minerals generally occur in the same rock but are rarely intermixed. Sericite, however, is absent in some specimens, and in one area on the east side of the pit some of the small dikes and plugs of granite porphyry contain abundant sericite but no hydrothermal biotite. On the whole, however, hydrothermal biotite is by far the more widespread and abundant. In the field the hydrothermal biotite-rich rock without sericite is difficult to distinguish from sericite-rich rock without hydrothermal biotite; consequently, they were not mapped separately. All the igneous-appearing rocks and a few of the sedimentary rocks in the pit have been affected to varying degrees.

The fine flaky dark mica occurring profusely throughout the Bingham mine as an alteration product is biotite, as determined by x-ray, but, since it is believed to have been formed under different conditions from the coarser biotite,

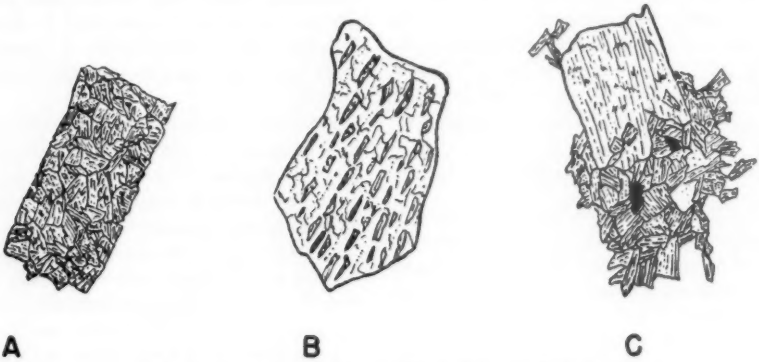

Figure 8.—Pseudomorphs of Hydrothermal Biotite

(A) former biotite crystal completely changed to variously oriented small hydrothermal biotite flakes. (B) former biotite crystal altered to quartz and hydrothermal biotite. (C) biotite crystal altered to hydrothermal biotite on lower end. Note the ragged and distended border of the hydrothermal biotite. Camera lucida drawing, ×32.

the term hydrothermal biotite is used. The only difference between the two is the size and form of the crystals.

The boundaries of many patches or clusters of randomly oriented hydrothermal biotite flakes have a shape resembling a biotite crystal (Fig. 8A). The edges of many of these pseudomorph patches are quite regular, but some are ragged (Fig. 8C). Occasionally, they are composed of biotite and quartz (Fig. 8B), and some pseudomorphs may contain as much as 80–90 per cent quartz.

The intensity of the development of sericite was not mapped because sericite could not be distinguished in hand specimen. However, the ease of recognizing hydrothermal biotite in the field made it possible to note two areas (not mapped) where it is profusely developed. In an area on the intermediate levels in Section 7, the rock was so rich in hydrothermal biotite (50–60 per cent) as to appear black, and, on the upper levels in Section 12, it is developed to a slightly greater degree than in the rest of the pit. The various rock types show some variations in the general effects that alteration Stage II had upon them; hence, each type is described.

Granite

Hydrothermal biotite is developed best in the granite. The small patches of concentrated flakes of hydrothermal biotite and its characteristic dark color make it fairly easy to detect in the hand specimen. The sericite of this stage, however, cannot be detected without the microscope. The dark mica flakes are found in clumps and clusters occasionally with ragged borders, disseminated throughout the rock; the size approximately equals the grain size of the original rock and may vary widely as the grain size of the granite varies. Most of the patches appear rather massive, presumably due to the very small size and random orientation of the small mica flakes. Though usually ragged in outline some of these patches which have been developed from pre-existing biotite may be quite regular. Where the patches are fairly large, they are blackish, but, where they are small, a brownish color is most prominent imparting a peculiar speckled salt and pepper appearance (Pl. 5, fig. 2). In some cases where the patches are fairly large, they seem to be interconnected and exhibit irregular and ragged borders. Some fairly coarse-textured dark rocks contain as much as 50 per cent hydrothermal biotite while most of the extremely fine-textured granite has a very small amount of it. No streaks, veinlets, or large areas of concentrated hydrothermal biotite are seen.

The sericite is confined principally to the feldspars and, where present in large amounts, colors them earthy white or tan. The kaolinite-illite development of Stage I also usually resulted in an earthy white or light brown color, and no field test was devised with which to distinguish confidently the presence of Stage I minerals from the sericite of Stage II.

The hydrothermal biotite is readily identified under the microscope by its prominent pleochroism, its high birerefringence (intense first-order colors up to second-order red), and the cedar grain or birds-eye structure characteristic of mica. The flakes, usually slightly larger than those of the sericite, are near .01 mm. to 0.1 mm. across, and various sizes may be found within one patch of hydrothermal biotite. It is found on or within all the minerals except quartz, and this includes the alteration minerals of Stage I (Pl. 3, fig. 5). In many instances the flakes cross mineral boundaries. Pseudomorphs of hydrothermal biotite after biotite are discovered with straight boundaries on all sides, but more often a portion of the boundaries are ragged (Fig. 8). Some are entirely ragged. Occasionally the brown mica flakes composing the rims of pseudomorphs are either coarser for finer-grained than those of the core. Very often fine-grained quartz accompanies the brown mica in the pseudomorphs, and in some instances as much as 90 per cent of the area may be quartz (Fig. 8B).

Up to half the original biotite in a specimen may be unaltered, and in many cases an unaltered biotite crystal is adjacent to thoroughly altered biotite. Most of the fine-grained dark mica areas are extremely irregular in outline; the individual flakes have no particular alignment or arrangement. These patches may be either isolated or interconnected, with individual flakes extending in irregularly shaped stringers from one area to another. The sizes of hydrothermal biotite flakes vary throughout the

rock, and there seems to be no relationship between intensity of alteration and the size of the mica flakes.

Sericite is found in practically all specimens, accompanying the hydrothermal biotite. It is distinguished microscopically by its lack of color, high birerefringence (intense first-order colors up to second-order red), flaky form, and its intermediate relief with the lowest index around 1 56. The sericite is found only in the feldspars and not intimately mixed with hydrothermal biotite. However, it is often present on or in a kaolinite-illite alteration area of Stage I. In the feldspars the sericite flakes are commonly spaced wide apart with unaltered feldspar between (Fig. 9); several characteristic arrangements of the flakes are seen. For instance, a "train" of randomly oriented sericite flakes crosses an otherwise unaltered feldspar crystal (Fig. 9A) as though the solution effecting the alteration had proceeded through a microfracture in the feldspar and the alteration has occurred only near the fracture. Several of these "trains" of sericite flakes may criss-cross within a single crystal. Many feldspars have altered only in part to sericite, while the rest of the feldspar crystal is clear and unaltered (Fig. 9B). The sericite flakes are frequently arranged with the long dimension of the shreds or flakes oriented in two directions forming a grid pattern over a feldspar crystal (Fig. 9C; Pl. 3, fig. 4). However, more commonly the sericite flakes have a completely random or feltlike arrangement with an even distribution throughout the crystal (Fig. 9D). With increasing intensity of alteration the amount of sericite increases, and the unaltered feldspar between the flakes decreases. In all the granite observed there was never more than 20 per cent of the rock changed to sericite, and except in a very few instances hydrothermal biotite always accompanied it.

Actinolite Syenite

The actinolite syenite was altered to a minor degree by Stage II type. The hydrothermal biotite and sericite are found in the same relationships and proportions and affecting the same minerals as in the granite but are not so abundant. Around the borders of the actinolite syenite areas where it is in contact with granite some actinolite and diopside have been altered to a fine flaky greenish mica rather than brown mica. The most intensely altered actinolite syenite contained about 10 per cent of Stage II alteration minerals.

Granite Porphyry

As in the granite, sericite is not detectable in hand specimen except for the alteration found in the small dikes and plugs of Section 5 and the upper part of Section 10 where hydrothermal biotite is essentially absent and a silky luster on intensely altered feldspar phenocrysts distinguishes the sericite. In the rest of the granite porphyry, hydrothermal biotite is abundant, but it generally amounts to about one third of that present in the granite and occurs principally as pseudomorphs after biotite. Only about one third to one half of the total biotite phenocrysts in the granite porphyry (which are readily observed in the hand specimen) are changed to pseudomorphic patches which generally have a regular outline but show a very dense structure.

Under the microscope about 90 per cent of the hydrothermal biotite in the granite porphyry is found in large patches with regular or irregular outlines and is believed to have developed largely from previously existing biotite. About 10 per cent of the hydrothermal biotite occurs in the quartz-orthoclase groundmass or in the feldspar phenocrysts distributed in small disconnected patches. Many of the pseudomorphs here contain considerable accompanying alteration quartz similar to that found in the granite. Some biotite phenocrysts have been bleached with the accompanying development of many oriented rutile needles. The sericite is found here in precisely the same manner as in the granite, being derived entirely from the feldspar and found in similar relationships (Figs. 9A, B, C, D).

Both second stage minerals are distributed throughout the granite porphyry. Where a later stage (Stage IV) has altered large areas in the granite porphyry, the hydrothermal biotite and sericite of Stage II have been obliterated, and identity of Stage II in these areas is therefor lost.

In aplite the alteration minerals of Stage II have been developed to about the same extent and intensity as within the granite porphyry proper.

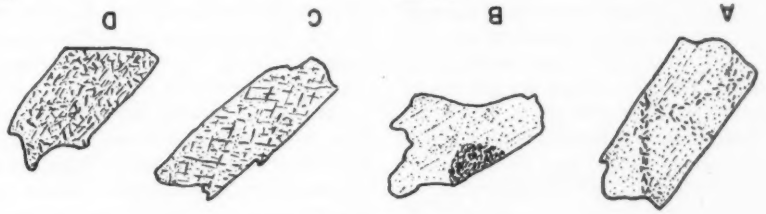

FIGURE 9.—SERICITE IN FELDSPAR

(A)—aligned in "trains." (B)—present only in part of a crystal. (C)—flakes oriented in a grid pattern. (D)—flakes in felt-like or random arrangement. Camera lucida drawing, ×26.

Biotite-Quartz Latite Porphyry

In practically all the dike rocks classified as biotite-quartz latite porphyry most of the feldspar phenocrysts have been completely changed to sericite, whereas hydrothermal biotite occurs principally on the fringes of the biotite and to a slight extent in the groundmass.

Granitized Quartzite

The typical granitized quartzite contains but a very little hydrothermal biotite, and no well-developed dark mica pseudomorphs were found. In only one or two specimens was hydrothermal biotite present in sufficient quantity to give the rock a brownish appearance. The smaller biotite flakes are more or less randomly oriented and disseminated. The sericite in the very fine feldspar aggregates is usually in such small flakes that very high magnification was necessary to resolve them. When observable, however, they are found to be present in an open disconnected manner similar to that found in the feldspar of the granite porphyry and granite.

Feldspar Network

In practically all cases quartzite having the feldspar network contains very fine-grained unmixed hydrothermal biotite and sericite both of which are present only in the feldspars. The flakes are generally quite small and randomly oriented, but occasionally some areas or patches are 0.5 mm. across and are believed to have been formerly occupied by feldspar. In some specimens the feldspar is partially altered only to sericite.

Sediments

Some hydrothermal biotite was found in a few unfeldspathized quartzites. It is recognized in hand specimen by the presence of some very small (0.5 mm.) widely spaced, brownish-looking spots disseminated throughout the rock.

In the limestones hydrothermal biotite is generally restricted to the edges of some of the xenoliths, or skialiths. In some places enough dark mica has been introduced to make up as much as 50 per cent of the rock, and occasionally the individual flakes become so large that cleavage units can be discerned megascopically.

Relations between Hydrothermal Biotite and Sericite

In many slides, hydrothermal biotite and sericite adjoin, and the boundary relations suggest that the dark mica is later than the sericite. However, there are no crosscutting veinlets or other visible structural features to show that the two minerals were developed at separate stages, and it is therefore believed that the two are closely related in time and environment of formation.

Character of Altering Solutions

It has been fairly well established (Gruner, 1944, p. 587) that true sericite (fine muscovite) probably forms in acid solutions above 350°C. and in alkaline solutions below 350°C. The field of formation of biotite is not so well known.

However, one might extrapolate that, since biotite is so similar to sericite, both might have approximately the same general environment of formation. There is no critical mineral accompanying this stage of alteration which would place the temperature of formation above or below 350°C. or the pH of the solution above or below 7. Therefore neither the pH nor the temperature of the solution can be determined.

Chemical analysis Number 2, Table 2, is of an intensely altered (50 per cent alteration minerals) granite containing hydrothermal biotite and sericite. This analysis is not radically different from an analysis of the comparatively unaltered rocks, Numbers 2 and 4, Table 1. The greatest difference is in the Fe_2O_3 content, which jumps from 1.88 per cent to 3.74 per cent in the altered rock. All other figures remain within a reasonable and expectable range considering the wide variation in composition of the original rock. The introduction of iron was no doubt responsible for the profuse formation of biotite. Analogous percentages for silica, alumina, magnesia, soda, and potash in the analyses indicate that not much if any of these materials were introduced.

Paragenesis

When Stage I minerals, kaolinite and illite, were found with Stage II minerals, the hydrothermal biotite and sericite frequently formed on the Stage I minerals, and occasionally veinlets of Stage II minerals traversed Stage I minerals (Pl. 3, fig. 5). This evidence is admittedly purely petrographic, but it is fairly conclusive that Stage II followed the development of illite and kaolinite of Stage I.

Hydrothermal Stage III—Chlorite-Hydrothermal Biotite

General Statement

Alteration Stage III, resulting in the development of profuse chlorite and hydrothermal biotite in granite, is readily recognizable in the field. The dark mica is of the same type as that found in Stage II though the individual flakes are generally much smaller. The term chlorite here is used to cover several mineral species of the chlorite group. At the beginning of the study seven chlorites—penninite, antigorite, clinochlore, delessite, diabantite, jenkinsite, and prochlorite—were distinguished optically. However, there seemed to be no spatial or areal relationship of these chlorites to one another, and in the present report it seemed proper to use the general term chlorite rather than any of the species names.

The granite which has been intensely altered to chlorite and hydrothermal biotite of Stage III appears very dense, nearly black, and when wet resembles black limestone. Where the alteration is less intense, the original granitoid texture may be discerned, though the granite is still quite dark. Alteration is best developed near the western margin of the pit (Sections 1 and 6) where it is readily seen as dark streaks. These streaks range in width from a few millimeters to slightly over 200 feet. Three irregular areas were mapped on the southern middle and upper levels (Sections 7 and 12), but the specimens in this region show less intense alteration than do the rocks to the west. Many additional small pods or patches of this alteration type are found everywhere within the granite area, but because of their small size they were not mapped.

Microscopic Features

Under the microscope, the hydrothermal biotite and chlorite always occur in a fine flaky form often intimately mixed. In the most intensely altered rocks no feldspar or biotite has been left unaltered, and the quartz units have been replaced principally on their edges. In less intensely altered rock some of the feldspar and original biotite may be unaffected. The alteration minerals occur either as very fine-grained felted irregular solid mats or, in less intensely altered rocks, as fine flakes scattered through feldspar and biotite.

Chlorite Orbicules

Of special note are some chlorite orbicules which have formed in granite as a result of Stage III alteration. In these orbicules the chlorite has formed with spherically curved cleavages, giving a distinct onionlike structure to the mass. The largest orbicule seen was 22

mm. in diameter (Pl. 5, fig. 1), but the average size is around 10 mm. with some as small as 2 mm. in diameter. The chlorite orbicules appear to have grown by replacing the minerals which surrounded it. It is not yet known why these orbicules have developed in such a unique manner.

Character of Altering Solutions

The minerals of the chlorite group have not been investigated extensively, and very little is known of the hydrothermal physico-chemical conditions under which they form. However, it is thought (T. S. Lovering and J. W. Gruner, Personal communication) that chlorite probably forms in neutral to slightly alkaline solutions at temperatures that range from near 100° to about 400° or 500°C. The intimate mixture of fine biotite and chlorite suggests they were formed at the same time and hence from the same type solutions. If biotite forms under similar conditions to sericite (alkaline solutions below 350°C.), then it may be argued that, since chlorite probably forms in alkaline or neutral solutions, the temperature at the time of Stage III alteration was probably lower than 350°C.

Chemical analysis Number 3, Table 2, was made of an altered rock of Stage III. The only significant chemical difference between this rock and the more or less unaltered granite (Table 1, Analyses 1, 2, 3, 4) is the expected higher magnesia and iron content. The percentages jump from a maximum of 2.7 for iron in unaltered rock to 5.1 and 4.1 for magnesia to 7.44 in the altered rock.

Paragenesis

The position of Stage III in the hydrothermal paragenetic sequence is very clear from structural evidence. In the field alteration streaks or bands of Stage III minerals cut through alteration Stage II, thus definitely dating the chlorite-hydrothermal biotite alteration.

Hydrothermal Stage IV—Quartz-Sericite

General Statement

The fourth alteration stage finds quartz and sericite extensively developed along variously oriented fractures in granite, actinolite syenite, granite porphyry, and granitized quartzite. The alteration has developed symmetrically on either side of the fractures producing bands easily recognized in the field. Where the bands are so wide and numerous as to coalesce leaving no unaltered rock between, as is common in the granite porphyry, the resultant rock somewhat resembles quartzite.

Field Features

The quartz-sericite alteration type clearly follows fractures in all cases. It colors the rock gray and converts it to a very dense chalcedonic-appearing material. Where the dark granite and actinolite syenite have been affected, these alteration areas appear as light streaks and bands (Pl. 5, fig. 5), but where they traverse the light-colored granite porphyry they appear darker than the host. It is difficult to recognize them in the granitized quartzite since the colors and textures are similar.

The alteration is always symmetrically developed on either side of a fracture, and the band or streak maintains a more or less constant width for a given fracture. The width of these bands ranges from 2 mm. to 10 cm. with most falling between 5 mm. to 3.5 cm., and the outer margins always appear as sharp contacts with the host rock. Along some outer borders in the granite a concentration of hydrothermal biotite formed, producing a darker streak on either side of the alteration area. Some fractures are open and empty, but most of them are filled either with glassy quartz of Stage V or sulfides of Stage VI, and frequently with both.

There seems to be no particular relation between the width of the alteration band and the width of the fracture. The fractures controlling the alteration are not preferentially oriented and are distributed entirely at random. They are usually fairly straight but may criss-cross one another at various angles in an intricate network. Because they are so easy to detect in the dark granite, some of the wider alteration bands (2–3 inches) have been traced in the field as far as two or three levels (200 or 300 feet).

Microscopic Features

Under the microscope the most intensely altered rocks of this stage are found to be com-

posed entirely of quartz and sericite in about equal amounts. The sericite, which occurs in irregular patches, is fine-grained, and the flakes are usually arranged in a feltlike manner, forming solid mats. The quartz is medium-grained with interlocking boundaries. Some of the alteration bands show less intense alteration, but this is also found to have quartz and sericite. Sericite is much less abundant and is found only in the position of former feldspars, biotite, and hydrothermal biotite. Some unaltered feldspar may remain. Where alteration has been less intense, the porphyritic texture of the granite porphyry is still readily determinable. Where an alteration band cuts large pink feldspar phenocrysts, the alteration seems to be predominantly sericite with sharp outer boundaries and contains quartz only near the center of the alteration band (Pl. 3, fig. 6).

There seems to be no particular gradation in the intensity of the alteration from the center of an alteration band to the outer margin; the whole band shows either intense alteration or moderate alteration. Although in most rocks the outer border of a band appears very sharp in hand specimen, under the microscope it is seen to be fairly gradational over a distance of 0.1 to 0.2 mm. Of special interest are bands of a dark mineral which sometimes occur just beyond the border of the alteration bands in granite (Pl. 4, fig. 6). Under the microscope this dark border is observed to be composed of hydrothermal biotite in greater concentration than in the granite, and within the alteration band itself are many bleached biotite crystals and hydrothermal biotite aggregates. This suggests that the altering solutions removed the iron and magnesia from the dark mica within the band to the position just beyond its border and there formed the dark mica. This condition may be somewhat similar to that commonly thought of as a "basic front."

Distribution

The alteration bands, though observed over the entire open pit, seem to be more concentrated in the western part of the mine. The amount of the altered rock generally ranges between 2 and 20 per cent of the whole, but, in many cases where alteration bands are wide and frequent, these figures may grade up to 100 per cent. The areas mapped as Stage IV on Plate 1 indicate the areas where alteration constitutes 25 per cent or more of the rock. Some fairly large areas of 100 per cent altered rock are found principally in the granite porphyry on the higher levels to the west in Sections 1, 2, 6, and 7. Only a few small patches of 100 per cent altered rock are found in the granite. The intensely altered specimens often closely resemble quartzite, and a thin section was necessary in many instances to determine the true nature of the rock.

Character of Altering Solutions

The solutions that accomplished the Stage IV alteration evidently obtained access into the rock along fractures and soaked symmetrically into the walls on either side of the fracture. The only significant chemical addition was SiO_2 since the material to form sericite is present in the feldspars and biotite. The solutions extracted MgO and FeO and perhaps a little K_2O from the biotite and hydrothermal biotite and either removed them or deposited them nearby as disseminated material or as dark mica streaks along the edges of the alteration band. No diagnostic hydrothermal mineral was developed during this alteration stage by which to estimate the temperature or pH of the solutions. It is believed that quartz may form under all types of physico-chemical environments and, as stated elsewhere, seicite is formed either in acid solutions above 350°C. or in alkaline solutions below 350°C. There is no evidence to indicate which condition existed at the time of the alteration.

Analysis numbers 4 and 5, Table 2, are of rock showing alteration Stage IV. Of significance is the fact that the SiO_2 has jumped from 60–68 per cent to approximately 74 per cent and that Fe_2O_3 and MgO have declined rather markedly. Analysis number 3, Table 1, is of a comparatively unaltered granite and was taken from rock approximately 3 mm. from an altered band. Analysis number 5, Table 2, is of the altered band. Note that SiO_2 rises from 66.71 per cent in unaltered rock to 73.61 per cent in the altered band and that the MgO and Fe_2O_3 and Na_2O significantly decline.

Paragenesis

The sequence of formation of Stage IV is very clear from field evidence. These alteration bands traverse or cut through the hydrothermal biotite-sericite rock of Stage III, the hydrothermal biotite-sericite rock of Stage II, and the kaolinite-illite rock of Stage I and are in turn cut by barren quartz veins of Stage V.

Hydrothermal Stage V—Quartz

The quartz of Stage V occurs in sharp-bordered veins which obviously filled fractures, and no minerals are altered as a result of its formation, hence stage V cannot be strictly called an alteration stage (Pl. 5, fig. 5). The quartz is clear to milk white and glassy and often contains vugs. The veins traverse all the rock types of the pit with random orientation and range in width from 0.5 mm. up to 10 cm. with the average near 1 to 7 mm. They are generally so widely spaced that no mapping of them was attempted. Sulfide crystals are found in many of the vugs.

As quartz is not diagnostic of temperature and pH of depositing solutions, no evidence as to the environment of formation of the Stage V quartz is possible. Study of the liquid inclusions in the quartz (not attempted here) could reveal information regarding the temperature of the quartz formation. In the field these veins cut through alteration bands of Stages III and IV (Pl. 5, fig. 5) and in turn are cut by sulfide veinlets, thus placing Stage V as definitely pre-sulfide and post-Stage IV.

Stage VI—Sulfides

General Statement

Although the presence of sulfides in the copper deposit at Bingham is the most important economic consideration, intensive study of the sulfide minerals was not attempted here for various reasons. This study was confined to the relationships of the stage of sulfide development to the overall hydrothermal history of the deposit. Evidence is conclusive that all the sulfides were formed during one clear-cut period in the hydrothermal sequence, but earlier work shows that the different sulfide minerals were probably deposited at different times within this period, and it is recognized that these features deserve special study. Beeson's (1916) early work on the sulfide minerals at Bingham shows a picture of the sulfide problem at that time, but perhaps a new study is now warranted because of new and deeper exposures. The large and extensive problem of oxidation and enrichment of the ore body, though also important, would not materially add to the present study and was therefore neglected.

Sulfide Minerals

Sulfide minerals believed to have been deposited by hydrothermal solutions have been determined as digenite (x-ray determination), chalcopyrite, bornite, enargite, molybdenite, and pyrite.

The digenite occurs in very small (0.1 mm. diameter) individual units disseminated throughout the deposit. It usually is accompanied by one or several other sulfides. Chalcopyrite is the most widespread ore mineral and occurs predominantly as small (1.0 mm. maximum diameter) units also disseminated throughout the rock, but occasionally small veinlets of it are noticed. In one instance, a veinlet of chalcopyrite, about 6 inches long and 2 cm. wide was found in granite. Chalcopyrite is also found in many instances as well-shaped crystals in vugs within the quartz veins of Stage V.

Most of the bornite is present in the lower levels of the pit. It occurs in a manner similar to the chalcopyrite—that is, as small disseminated grains and rarely as small and disconnected veinlets.

Enargite is rare in the mine but, where found, is present in the vugs of the Stage V quartz.

Molybdenite occurs in very small units distributed throughout all the rocks in the mine; it is almost universally associated with the copper minerals. The largest molybdenite crystals are obtained, however, where it has crystallized in vugs in the quartz of Stage V.

Pyrite is by far the most abundant sulfide mineral in the mine. It forms small crystals disseminated throughout the rock, small veinlets, and very often well-developed crystals in vugs in the quartz of Stage V (Pl. 5, fig. 5).

Pyrite is also abundant in the sediments around the periphery of the igneous-looking rocks and in some cases may replace fairly large masses, particularly in the limestone.

Distribution

Pyrite is found in small crystals or large replacement bodies in all types of rock within the pit. Enargite occurs only in vugs of Stage V quartz while chalcopyrite, pyrite, bornite, and digenite are found in varying amounts in all the igneous-looking rocks and to a minor extent in some of the quartzites and limestones near the granite and granite porphyry.

The copper, iron, and molybdenum sulfide minerals do not seem to favor in occurrence any special alteration type or fundamental rock type within the mine except that they are generally more sparse in quartzite and limestone.

The sulfide minerals disseminated in the igneous-appearing rocks seem to favor positions within or adjacent to biotite (Schwartz, 1947, p. 326).

Alteration

The sulfide minerals in the pit were formed by (1) replacement in the solid rock, and (2) precipitation in the vugs of the quartz of Stage V. No rock-alteration effects were found which could be related to the development of the sulfide minerals.

Nature of Depositing Solutions

Available data indicate that none of the minerals formed during the sulfide stage are diagnostic of the physico-chemical nature of the depositing solutions. The question is raised as to just how the sulfide minerals were disseminated so uniformly throughout the rocks at Bingham. Many instances are found where quartz of Stage V containing large vugs may have only a few sulfide crystals deposited on the quartz, whereas, in the solid rock adjacent to the quartz vein, disseminated units of sulfide minerals in the usual concentration are present (Pl. 5, fig. 5). If the sulfide-bearing solutions traveled through the rock in a soaking manner, the connected vugs in the quartz veins would probably act as channels, accelerating the movement of the solutions. It might be expected then that the vugs, having a more constant and fresh supply of sulfide-depositing solutions, ought to be filled with sulfides. This is true only rarely, and the vast majority of quartz veins which show vugs contain but a small amount of sulfides. When the overall spatial distribution of the sulfides in the rock is observed, the sulfide minerals are about equally spaced whether there is a vein with vugs in the area or not.

From this it is suggested that perhaps a new explanation of the mechanism of the sulfide introduction and deposition might be forthcoming. The sulfide-bearing solutions could have penetrated the rocks at Bingham and more or less come to rest without any deposition. This penetration when completed would soak not only through the pores of the rocks but also through the vugs in the quartz. Then, a more or less uniform concentration of sulfide material would be accomplished through diffusion throughout the soaked-up rock. At some time when conditions were right, crystallization of sulfides began with nucleation at more or less equally spaced points in the rock whether in a vug or not. The nuclei then grew, being fed by diffusion from the solutions in the immediate neighborhood of the growing crystal. Hence it might be thought that each crystal or unit represents the deposition of all the sulfide material from the solutions which were present just before precipitation. This would necessarily imply a rather concentrated ore solution. For example, if the total porosity of the rock (and hence volume of the ore solution) is assumed to be 10 per cent, this formed 1 per cent copper sulfide ore by this process, the rock density is taken as 2.6. The solution would have to contain 0.26 gm each of sulfur and copper per cc—*i.e.*, it would contain 26 weight per cent each of Cu and S if it had a density of 1.0. At this stage, local cracks may have formed resulting in the streaming in of sulfide-bearing solutions forming occasional small solid veinlets of sulfides. This whole process could thus be thought of not as a dynamic one but rather a comparatively stagnant soaking action.

There is a possibility that the uniform disseminated distribution of sulfides may be the result of an original uniform disseminated distribution of magnetite in the original igneous

rock and that the magnetite crystals, when the sulfide-bearing solutions traversed the rock, simply formed nuclei for sulfide replacement (Boutwell, 1905, p. 168). The present position of each crystal of sulfide then would largely be in the position of a former magnetite crystal. Magnetite was discovered in the limestones of the upper middle levels on the south side, but, in all the sections of the ingeous-appearing rock examined, not a single magnetite unit was found. If the above process is correct, some magnetite crystals should have been left unreplaced.

Hydrothermal Stage VII—Sericite-Allophane

Following the deposition of the sulfides, sericite and allophane were formed in cracks and fractures. The two minerals have not yet been found together. The sericite occurs in veins up to an inch wide, principally on the east side of the pit. It is dense, white, with a satiny appearance and a very soapy feel. The allophane is quite widespread. One area near the middle levels on the south side contains so much allophane that the rocks appear white from a distance. The allophane is white, dense, and porcelainlike and forms in veins, the same as the sericite. Both minerals have been deposited on top of sulfide minerals, thus establishing their age as post-sulfide.

Summary of the Hydrothermal Alteration and Mineralization Stages

In the first alteration stage following the formation of the rocks, kaolinite and illite were formed as an alteration product from feldspars in an area more or less confined to the north-central part of the pit. The temperature of the depositing solutions was probably below 350°C, and the solutions were acid. No appreciable amount of material was added to or extracted from the rock as a result of this stage.

Stage II finds sericite and hydrothermal biotite extensively developed in all the igneous-appearing rocks in the pit. Petrographic evidence separates Stage I from Stage II; microscopic veinlets of sericite and hydrothermal biotite are found in kaolinite and illite. Iron and magnesia were introduced by the solutions, but not enough is known regarding the field of formation of hydrothermal biotite or sericite to state confidently the temperature and pH of the solutions. On the basis of the presence of sericite, the solutions either were at temperatures above 350°C and acid or below 350°C and alkaline.

Stage III is found principally in the granite as dark streaks and bands consisting of an intensive development of chlorite and hydrothermal biotite. These streaks cutting the granite establish this alteration as post-Stage II. Chlorite is believed to form in solutions which are neutral to slightly alkaline, and the hydrothermal biotite, so closely associated with the chlorite, probably forms in neutral to alkaline solutions which are below 350°C. Therefore, the alteration Stage III might have been accomplished by slightly alkaline solutions below 350°C.

The quartz and sericite bands of Stage IV cut Stage III alteration as well as Stages I and II, thus establishing its age. Sericite or quartz are not diagnostic of any particular environment, and nothing can be said regarding the temperature and pH of the solutions. Silica was introduced in profuse amounts.

Stage V saw the development of quartz only in veins and is not an alteration stage. In the field the veins cut the quartz sericite alteration of Stage IV as well as all earlier alteration stages, and it is therefore later. Quartz is believed to form in all types of solutions at all temperatures, hence no statement can be made regarding the temperature and pH of the solutions. Silica was the only substance introduced at this stage.

Sulfide crystals of Stage VI are in vugs in the quartz of Stage V, and small sulfide veinlets cut the quartz veins. Thus the sulfide stage is separated from the quartz mineralization of Stage V. Since various sulfides may be formed in all types of solutions in all ranges of temperatures, no interpolation can be made regarding the temperature and pH of the solutions. Copper, iron, sulfur, molybdenum, a little arsenic, silver, and gold were introduced by these solutions.

The minerals of the last hydrothermal stage, allophane and sericite, coat sulfide crystals, thus establishing that the allophane-sericite

stage is post-sulfide. Neither sericite nor allophane is diagnostic of any hydrothermal environment of formation, and the temperature and pH of these solutions remain unknown. Potash, alumina, silica, and H_2O were introduced.

The solutions of the first four stages definitely altered some of the minerals in the rocks, and the solutions of Stages I and II soaked through the rock more or less uniformly. The solutions of the third and fourth stages traversed the rock through fractures and altered the rock adjacent to them. The solutions of the last three stages introduced material, and no alteration in the above sense was effected. Solutions of the fifth and seventh stages followed fractures, while, in the sixth or sulfide stage, solutions penetrated the rock uniformly, suggesting a soaking action.

Origin of Solutions

The problem rises as to the origin of the post-magmatic or post-granitization solutions and the localization of these solutions within the Bingham copper mine area. The solutions evidently came from depth, being derived as a differentiate of the original magma or the neomagma, which ever view is taken. The broader geographic localization may be due to the presence of a deep-seated channel in the vicinity of Bingham through which the solutions moved from a fairly deep source. The local position of the alteration stages and introduced minerals in the rock is probably the result of: (1) the original higher porosity of the igneous-appearing rock plus the intense fracturing which was localized in these same rocks and occurred concurrently with the entrance of at least the later solutions; (2) the original composition of the rocks for some reason caused the deposition of minerals from the solutions to favor the igneous-appearing rocks. The specific distribution of alteration types among the rock types is thought to be more a matter of spatial coincidence than a result of specific chemical "affinities".

It is not within the province of this study to speculate on the separation of the solutions from the magma or to offer reasons for the several distinct stages of mineralization; these problems will be left to others. Perhaps when the entire Bingham or whole Oquirrh Mountain area has been intensively studied, some of the problems along these lines will be solved satisfactorily.

General Conclusions

The studies of the hydrothermal alteration at Bingham have resulted in the recognition of seven distinct mineralization periods. From this information interpretations have been made regarding the temperature and pH of the altering solutions of Stages I and III. Perhaps a reasonable and acceptable hypothesis could be made regarding the pH and temperature of Stages II and IV, but similar information regarding the last three stages would be purely guesswork.

The geologic and mineralization features of the Bingham disseminated copper deposit could be compared with other similar deposits, but no new information would be forthcoming since an adequate comparison has already been made by Schwartz (1947, p. 323–346), Kerr (1951, p. 478–480), Anderson (1950, p. 621–625), and others. A review of the hydrothermal studies of porphyry copper deposits by Kerr (1951) and Kerr *et al.* (1950) at Silver Bell and Santa Rita; Anderson (1950) at Bagdad; Gilluly (1946) at Ajo; Peterson *et al.* (1946) at Castle Dome, and Schwartz (1947) at San Manuel shows perhaps as many dissimilarities as similarities to the hydrothermal features at Bingham.

One may take the similarities of the group such as the argillic (illite and kaolinite at Bingham), biotitic, sericitic, and silicic alteration and propose that these features will likely be present to some extent in all "porphyry copper" deposits and that they might be used confidently in further exploration. At Bingham, however, the ore and protore mineralization extends practically to the full limit of the altered area making the alteration target there essentially no larger than the ore target; therefore the hydrothermal alteration is of little practical use as a guide to the ore finder at Bingham. This condition may not be true for all deposits, and the characteristic alteration above mentioned may be of great value to the ore finder in other areas.

The granitization and the feldspar network halo associated with the Bingham ore body possibly offers a new or different line of thought to the student of porphyry copper deposits.

There is a possiblity that some of the dissimilarities of the various deposits mentioned above may be more apparent than real, due to the differences in background of the different workers and the time and intensity of the studies.

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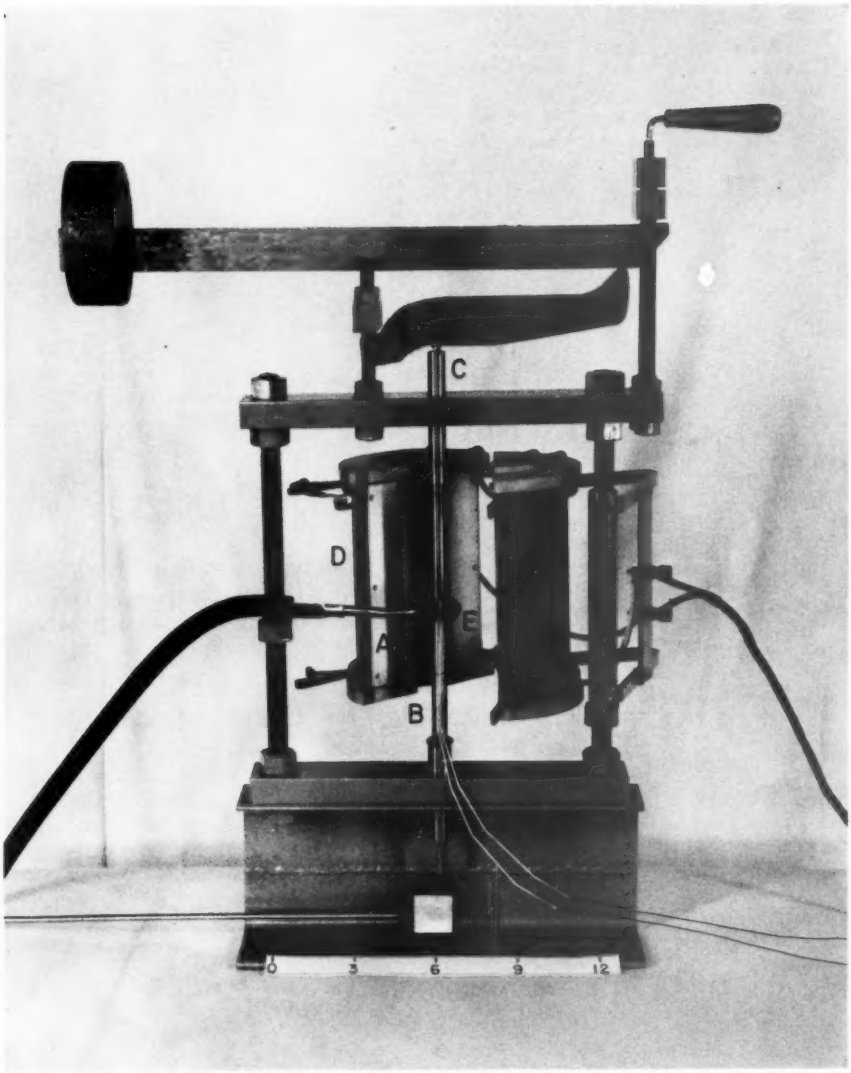

PHOTOGRAPH OF ONE MODIFICATION OF THE IMPROVED HYDROTHERMAL QUENCHING APPARATUS

A — Sample holder; B — Lower stem; C — Weighted piston; D — Furnace; E — Air blast. Scale in inches.

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Short Notes

IMPROVED HYDROTHERMAL QUENCHING APPARATUS

By John Van den Heurk

Introduction

A considerable amount of information has been acquired on the effect of temperature on anhydrous combinations of the principal rock-forming oxides. Data on these "dry" silicate mixtures have contributed to our understanding of the olivines, pyroxenes, and feldspars, and of the mutual-stability relations of many of the simpler rock-forming minerals. Systematic studies of the phase relations in mineral systems with water as a component have been few, but progress has been made on some of the micas, amphiboles, and zeolites. In many cases, the addition of water to a system has aided the growth of anhydrous minerals produced with difficulty in the dry way. Great impetus was given to these hydrothermal studies by the development of an apparatus by Dr. O. F. Tuttle (1948) which permitted the worker to expand greatly his range of experimental conditions and to expedite his procedures. During the last 5 years, as more units were put into operation, many improvements have been made on the apparatus, mainly in the direction of simple construction and low cost. More recently, students have expressed the desire to investigate hydrothermal systems and have been obliged to manufacture their own equipment. Primarily for this reason changes were made to simplify the construction of Tuttle's original apparatus. As a result, the number of man hours required to produce the apparatus has been reduced, and the cost lowered accordingly.

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Description of Apparatus

A photograph of one modification of the apparatus is given in Plate 1. The charge is held in a small pressure vessel (A), the lower end of which is coned. The pressure vessel is supported in a cone seat in the lower stem (B) through which the volatile materials are introduced. The cone-in-cone closure is maintained by a weighted piston (C) that bears on the pressure vessel. The weights, acting through the levers on the piston, counterbalance the pressure exerted by the volatile material. A furnace (D) surrounds the pressure vessel, and a pump delivers the required pressure through the lower stem. In this way a charge can be held at a fixed temperature and pressure. The products produced under these conditions are quenched by means of an air blast (E) directed at all sides of the pressure vessel. When the charge is sufficiently cool the pressure is released, and it is removed for study. Each of the essential components of the apparatus will now be described in detail.

The upper and lower cross bars are retained in parallel horizontal position by pairs of nuts equally spaced on two vertical threaded bolt rods ¾ inch in diameter. The bolt rod and bar may be obtained from a hardware store in lengths of several feet and cut with a hack saw to the necessary length. In order that the bolt rod holes as well as the piston and lower stem holes will be drilled in line, the two bars are laid on top of each other for drilling. The remaining four holes in the upper bar are also

drilled in the standard drill press. The lower cross bar may be bolted directly to a table top or to a beam as shown in the photograph. As many as six units have been mounted on a single raised beam. This arrangement permits the installation of valves in a convenient position directly under the inlet stem.

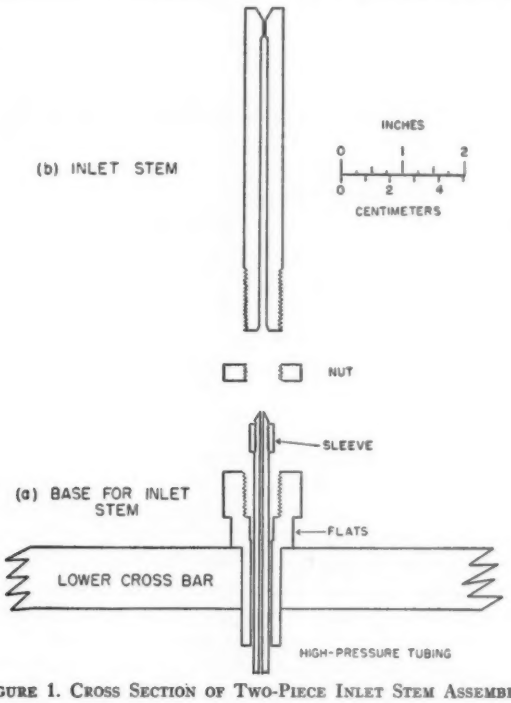

FIGURE 1. CROSS SECTION OF TWO-PIECE INLET STEM ASSEMBLY

The inlet stem is made in two pieces for ease of machining and to reduce the cost when expensive high-temperature alloys are employed. The lower part, (a) in Figure 1, is machined from a ¾-inch round of stainless steel (S.A.E. 18-8). The interior thread is more easily made by cutting on the lathe, but a novice may prefer using a ⅝-inch diameter tap having the standard 18 threads per inch. Although the flats can be milled, a satisfactory job may be obtained quickly by use of a tool grinder. The coaxial hole is of such size as to pass freely the commercially available high-pressure tubing. The seat is of a diameter to receive easily the retaining sleeve, an essential part of the cone-in-cone seal. If capillary high-pressure tubing is preferred, a one-piece cone and sleeve may be silver-soldered onto the tubing.

The upper part of the inlet stem, (b) in Figure 1, may be machined from a ⅝-inch round of stainless steel or a high-temperature alloy such as Haynes Stellite 25. The ends are coned 60° first, and then the 1/32-inch capillary hole is drilled ¼ inch deep from the bottom of the cone. The small hole is necessary to prevent convective loss of material from the sample. As the drilling of this hole is the most difficult operation in the construction, considerable patience is required. Greatest success has been achieved using a gun-type, carboloy-tipped drill. The drill is mounted in a pin vise having a hollow handle, which rests on a rod held in the tail stock chuck of the lathe. The drill is withdrawn after every half dozen or so revolutions of the piece, *all* chips are blown from the hole, and the drill is oiled. If attempts are made to drill too deeply at each insertion, the chips

bind the drill, and it breaks. When this happens, the piece must usually be abandoned.

The ⅛-inch hole in the upper part of the inlet stem is drilled 4 inches deep in a similar fashion. Care should be taken as the larger hole nears the capillary hole because the drill is particularly apt to snap as it makes junction with the capillary hole. The 4-inch-long hole can be drilled in a few hours by a beginner. The thread must be cut on the lathe for coaxial alignment. The thin nut on the upper part of the inlet stem acts as a support and rests on the face of the lower part of the inlet stem. This nut bears some of the excess load on the sample holder.

The sample holder may be machined from a ⅝-inch round of stainless steel or high-temperature alloy. The off-center $\frac{3}{32}$-inch thermocouple hole is drilled in a drill press ½ inch deep in the end of a round cut to ⅞ inch in length. The ⅛-inch hole is then drilled to a depth of $\frac{9}{16}$ inch from the other end, and the sample holder is coned to 59° as illustrated (Fig. 2). It has been the practice in this laboratory to recone the sample holder after each run, but this is not essential for runs at low temperatures. When the sample holder or the inlet stem has suffered deformation, it is wise to redrill the holes and recone to insure an adequate seal on succeeding runs. It is convenient to keep a large supply of sample holders on hand.

The sample may be held in an open crucible, a sealed tube, or a buffered capsule. It has been found useful to use a small pointed platinum rod as a pedestal for the charge. Such a pedestal, in addition to cutting down convective circulation, also keeps the charge close to the thermocouple.

The piston is simply a ⅝-inch round skinned down a few thousandths of an inch at the end extending into the furnace for clearance in case of bulging or warping. An axial groove is cut in the same end by means of a hack saw or cut-off wheel to make room for the thermocouple. When runs are made at high temperatures, the portion of the piston which extends into the furnace may be made of a high-temperature alloy. A press fit stem-in-cup union between the two pieces is quite adequate.

In order to maintain the cone-in-cone seal, a force greater than that exerted by the volatile must be imposed on the piston. This may be accomplished by a simple lever approximately 3½ feet long. However, a double lever, such as

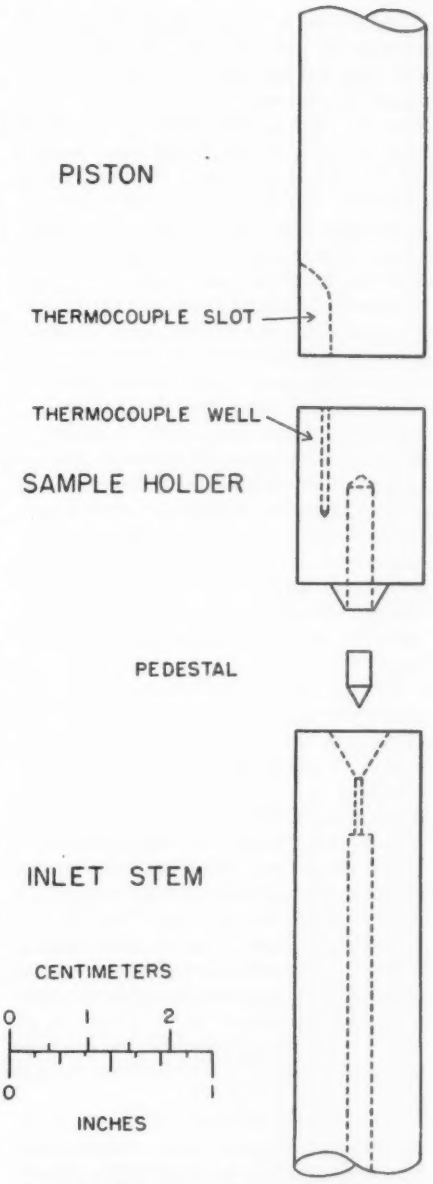

FIGURE 2. DRAWING OF SAMPLE HOLDER AND PORTIONS OF THE PISTON AND INLET STEM

shown in Plate 1, requires much less space and provides the same leverage. The bars are ⅜ inch thick and 1½ inches wide, and are easily shaped by trimming with a metal band saw or hack saw. The fulcrum points of the double lever are made of bars sliding freely between two ½-inch bolt rod uprights. The first fulcrum is adjusted by nuts on the uprights and need be changed only when a piston of different length is inserted. However, the second fulcrum must be adjusted for each run depending upon the size of the sample holder and on the temperature which expands the piston. For maximum efficiency the lever arms should be maintained in a horizontal position throughout the run. A load of approximately 13 pounds is required to balance a pressure of 30,000 psi using the lever system illustrated in Plate 1. In order to keep deformation of the cone-in-cone seal at a minimum, no more than the necessary weight should be used. At lower pressures, the weight may be moved along the upper bar toward the fulcrum to reduce the force exerted. Pieces of ¼-inch round may be brazed onto the levers at the bearing points for smooth operation.

A split combustion furnace, wound with nichrome wire, is satisfactory up to 1100°C. Furnaces of this type with replaceable heating units are commercially available. A Pt10Rh winding gives better service for long runs when higher temperatures are required. The furnace is centered by means of a split transite disc fastened to the top containing a hole the same size as the piston. The mounting is made of two soft iron bars which are easily bent into the desired shape. These are brazed onto a piece of ordinary ¾-inch pipe which slides easily on one of the bolt rod uprights of the frame. The region of highest temperature in the furnace is adjusted to coincide with the height of the sample holder by means of a nut on which the pipe mounting rests.

Temperature is adjusted by a resistor, variac, or other regulating device in the furnace circuit. The temperature is measured with a thermocouple inserted into the wall of the sample holder, the base of which is only ⅛ inch from the center of the charge.

The volatile material is pumped into the inlet stem through standard high-pressure tubing. An ordinary hydraulic automobile jack can be converted into a suitable hand pump for low-pressure work. An electrically driven pump may be controlled with a switch activated by the needle on the pressure gauge.

The products produced at the fixed pressure and temperature are quenched by opening the furnace and directing a stream of air at all sides of the sample holder. The air blast is effected by means of a copper tube, perforated with holes and bent into a U shape. A simple clip-on bracket of brass strip is soft-soldered to the tubing. The height of the air blast may also be adjusted by a nut on the bolt rod upright of the frame. When not in use, the air blast is swung aside. When the temperature has dropped sufficiently to quench the charge, the pressure is released by breaking a seal in the high-pressure line or by opening a valve. Only the weight, lever arms, piston, and sample holder need be removed. In order to preserve the structure on which its strength depends, it is desirable to quench the piston by plunging it into cold water. Such a quenching procedure would also lengthen the life of the inlet stem, but the time required to remove it prevents an adequate quench.

The apparatus described, with pistons and inlet stems of Stellite 25, has been operated up to 1050°C at 15,000 psi for runs as long as 1 hour. At lower temperatures the apparatus has performed well up to 45,000 psi and at lower pressures up to 1100°C. Argon, CO_2, and oil, as well as water, have been used successfully as pressure media, but it is likely that other volatiles may be used in the apparatus.

Conclusion

The improved hydrothermal quenching apparatus can be constructed in less than 100 hours at a cost of approximately $60.00 for materials (excluding furnace and pump) at the prevailing prices. Students without previous mechanical experience were able to construct the apparatus without assistance. It is hoped that the small size, low cost, simple construction, and ease of operation of this apparatus will promote the experimental investigation of volatiles in petrological problems.

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FOSSILS IN METAMORPHIC ROCKS

BY ARTHUR J. BOUCOT

Interesting but controversial points concerning the geology of New England are the determination of the age of the metamorphosed rocks, as well as the determination of the time or times when the metamorphism took place. The determination of the age of these metamorphic rocks is of great importance in reconstructing the Paleozoic paleogeography of the northern Appalachians. Dr. Walter Bucher's (1953) recent stimulating paper on *Fossils in metamorphic rocks* was therefore of special interest in summarizing present knowledge of the metamorphic fossils of New England[1] and elsewhere. To record and to scrutinize carefully any object from this metamorphic terrain that is purported to be of organic origin is most important. Critical comment on the nature of the specimen illustrated and identified as *Spirifer murchisoni* on Plate 1, Figure 7, of Doctor Bucher's paper should therefore not be taken amiss.

The specimen in question was recently examined, through the courtesy of Professor Charles G. Doll, by G. Arthur Cooper, Preston E. Cloud, Jr., Edwin Kirk, and the writer. Doctor Cooper was of the opinion that the specimen was not of organic origin, basing his conclusion on the relatively undistorted nature of the plications occurring as they do in a metamorphic rock in a highly deformed area. Doctor Cooper also felt that, admitting the possibility of the specimen being a fossil, it could not be assigned to a phylum, much less identified generically or specifically. He also pointed out that the plications are flatter than those of either *"Spirifer" macropleura* or *"Spirifer" murchisoni* and that the furrows, which are wider than the plications, are unlike those of the aforementioned species. Neither *"Spirifer" murchisoni* or *"Spirifer" macropleura* possesses bifurcating plications as does this specimen. Doctor Cloud felt that the specimen might be a brachiopod and possibly even a spiriferoid but that a generic identification was unwarranted. Doctor Kirk is of the opinion that the specimen is probably a fossil, but he would be unwilling to assign it to a phylum, much less to identify it generically or specifically. The writer feels that, if the specimen is a fossil, it might be a brachiopod. If it is a brachiopod, it might be a spiriferoid or any other form with similar ornamentation, as for example the Middle Ordovician to lower Silurian orthoid *Platystrophia*. However, no spiriferoid or *Platystrophia* with the same type of ornamentation as this specimen is known to Doctor Cooper. The writer feels that it is unjustified and misleading to label this disputed specimen as *Spirifer murchisoni*. By so doing the strata in which it was found at South Strafford, Vermont, would be dated Lower Devonian, an age determination not warranted on the basis of this specimen. Evidence afforded by fossils from metamorphic rock should not be pushed further than is scrupulously justified by the specimen in hand, and dubious determinations should not be allowed to stand unchallenged in the published record.

[1] Bucher omits perhaps deliberately the *"Cordaites* leaves," "small leaflet of *Sphenopteris*," and the *"Lepidodendron"* mentioned by David White in Emerson (1917) as occurring in the Worcester phyllite of Massachusetts, as well as the "graptolites" reported by Richardson (1908) from the metamorphosed Waits River formation of Vermont. The Worcester phyllite specimens seem to be lost, and the Waits River "graptolites" highly doubtful.

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FOSSILS IN METAMORPHIC ROCKS: A REPLY

BY W. H. BUCHER

I deserve censure for having used the name *Spirifer murchisoni* in the legend to Plate 1 without indicating that legitimate doubt exists concerning the specific identification which Doll had merely suggested as a possibility. To correct this error, the reader should put one question mark after *Spirifer* and two after *murchisoni*. But I see no reason to doubt that the specimen is the impression of a rather large spiriferoid brachiopod. The radiating pattern of costae widening toward the anterior margin with its characteristic crowding of the growth lines cannot well be the ghost of totally vanished crystals (of which no trace exists elsewhere in the rock) nor a mechanical trick, a *lusus naturae*.

When years ago Doll showed me the specimen, I looked it over carefully. Having seen many deformed fossils in nonmetamorphic rocks and recognizing the abnormal curvature of some of the ribs, it never occurred to me that the plications on this specimen were "bifurcating". Observations on deformed spiriferoid shells show that "squashing" readily changes the shape and crowding of the costae and can even create narrow secondary plications where they do not belong. In these features Doll's specimen does not go beyond the changes known from other deformed fossils in unmetamorphosed rocks.

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